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RESULTS

AMERICAN SCIENCE SERIES—ADVANCED COURSE

GEOLOGY

BY

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IN THREE VOLUMES

VOL. I.—GEOLOGIC PROCESSES AND THEIR RESULTS

SECOND EDITION, REVISED

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PREFACE.

In the preparation of this work it has been the purpose of the authors to present an outline of the salient features of geology, as now developed, encumbered as little as possible by technicalities and details whose bearings on the general theme are unimportant. In common with most writers of text-books on geology, the authors believe that the subject is best approached by a study of the forces and processes now in operation, and of the results which these forces and processes are now bringing about. Such study necessarily involves a consideration of the principles which govern the activities of geologic agencies. These topics are presented in Volume I, and prepare the way for the study of the history of past ages, which is outlined in Volume II.

The general plan of the work has been determined by the experience of the authors as instructors. Little emphasis is laid on the commonly recognized subdivisions of the science, such as *dynamic geology*, *stratigraphic geology*, *physiographic geology*, etc. The treatment proceeds rather from the point of view that the science is a unit, that its one theme is *the history of the earth*, and that the discussions of dynamic geology, physiographic geology, etc., apart from their historical bearing, lose much of their significance and interest. The effort has, therefore, been to emphasize the historical element, even in the discussion of special themes, such as the work of rivers, the work of snow and ice, and the origin and descent of rocks. This does not mean that phases of geology other than historical have been neglected, but it means that an effort has been made to give a historical cast to all phases of the subject, so far as the topics permit.

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Throughout the work the central purpose has been not merely to set forth the present status of knowledge, but to present it in such a way that the student will be introduced to the methods and spirit of the science, led to a sympathetic interest in its progress, and prepared to receive intelligently, and to welcome cordially, its future advances. Where practicable, the text has been so shaped that the student may follow the steps which have led to present conclusions. To this end the working methods of the practical geologist have been implied as frequently as practicable. To this end also there has been frankness of statement relative to the limitations of knowledge and the uncertainty of many tentative conclusions. In these and in other respects, the purpose has been to take the student into the fraternity of geologists, and to reveal to him the true state of the development of the science, giving an accurate and proportionate view of the positive knowledge attained, of the problems yet unsolved, or but partially solved, and of solutions still to be attained.

The theoretical and interpretative elements which enter into the general conceptions of geology have been freely used, because they are regarded as an essential part of the evolution of the science, because they often help to clear and complete conceptions, and because they stimulate thought. The aim has been, however, to characterize hypothetical elements as such, and to avoid confusing the interpretations based on hypothesis, with the statements of fact and established doctrines. Especial care has been taken to recognize the uncertain nature of prevalent interpretations when they are dependent on unverified hypotheses, especially if this dependence is likely to be overlooked. If this shall seem to give prominence to the hypothetical element, it should also be regarded as giving so much the more emphasis to that which is really trustworthy, in that it sets forth more frankly that which is doubtful. Hypothetical and unsolved problems have been treated, so far as practicable, on the multiple basis; that is, alternative hypotheses and alternative interpretations are frequently presented where knowledge does not warrant positive conclusions.

In many cases the topics discussed will be found to be presented in ways differing widely from those which have become familiar. In some cases, fundamentally new conceptions of familiar subjects are involved; in others, topics not usually discussed in text-books are stated with some fullness; and in still others, the emphasis is laid on points which have not commonly been brought into prominence. Whether the authors have been wise in departing to this extent from beaten paths, the users of the volumes must decide.

The work is intended primarily for mature students, and is designed to furnish the basis for a year's work in the later part of the college course. By judicious selection of material to be presented and omitted, the volumes will be found useful for briefer courses, and by the use of the numerous references to the fuller discussions of special treatises, they may be made the basis for more extended courses than are commonly given in undergraduate work. The attempt has also been made to make the volumes readable, in the belief that many persons not in colleges or universities will be interested in following a connected account of the earth's history, and of the means by which that history is recorded and read. Antecedent elementary courses in geology will not be necessary to the use of these volumes, though such courses may be helpful.

The arrangement of themes adopted is such as to bring to the fore processes with which all students are immediately in contact, and which are available for study at all seats of learning. The commoner geologic agents, such as the atmosphere and running water, have been elaborated somewhat more fully than is customary, and the common rather than the exceptional phases of the work of these agents have been emphasized, both because of their greater importance and their universal availability. The text has been so shaped as to suggest field work in connection with these topics especially, since work of this sort is everywhere possible.

After the preliminary outline, which is intended to give some idea of the scope of the science, and of its salient features, and to show the relations of the special subjects which follow, the order of treatment is such as to pass from the commoner and more readily apprehended portions of the subject to those which are less readily accessible and more obscure. Following the same general conception, the treatment of the topics is somewhat graded, the earlier chapters being developed with greater simplicity and fullness, while the later are somewhat more condensed.

Many acknowledgments are made in the text and foot-notes, but it is impossible to adequately acknowledge all the sources which have been drawn upon, since the whole body of literature has been laid under contribution. The authors especially acknowledge the generous assistance of Professor J. P. Iddings in connection with the chapter on The Origin and Descent of Rocks; of Dr. F. R. Moulton, Professor C. S. Slichter, Professor L. M. Hoskins, Mr. A. C. Lunn, and Mr. W. H. Emmons in connection with mathematical problems; of Professor C. R.

Barnes in connection with the geologic functions of life; and of Professor Julius Stieglitz in connection with chemical subjects.

The illustrations have been selected from numerous sources, which are usually acknowledged in the text. Especial acknowledgment is due to the U. S. Geological Survey for the use of numerous photographs and maps, and to Mr. G. A. Johnson, who has made many of the drawings reproduced in Volume I. The authors are under even larger obligations for assistance in the preparation of Volume II, for which acknowledgment will be made in the proper place.

UNIVERSITY OF CHICAGO, January, 1904.

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GEOLOGY

CHAPTER I.

PRELIMINARY OUTLINE.

GEOLOGY treats of the structure of the earth, of the various stages through which it has passed, and of the living beings that have dwelt upon it, together with the agencies and processes involved in the changes it has undergone. *Geology is essentially a history of the earth and its inhabitants.* It is one of the broadest of the sciences, and brings under consideration certain phases of nearly all the other sciences, particularly those of astronomy, physics, chemistry, zoology, and botany. It also embraces the earlier expressions of mental development and of life-relationships, chiefly as found in the lower animals.

Subdivisions.—Naturally so broad a science has many special aspects which constitute subdivisions, in a sense, though they are rather dominant phases than independent sections. That phase which treats of the outer relations of the earth is *Cosmic* or *Astronomic geology;* that which treats of the constituent parts of the earth and its material is *Geognosy,* of which the most important branch is *Petrology,* the science of rocks. That branch which investigates the structural arrangement of the material, or "the architecture of the earth," is *Geotectonic,* or *Structural geology;* while that which deals with the surface changes and topographic forms, that is, with the face of the earth, is *Physiographic geology.* The study of the fossils that have been preserved in the rocks, and of the faunas and floras that these imply, constitutes *Paleontologic geology,* or *Paleontology.* The treatment of the succession of events forms *Historical geology.* This is chiefly worked out by the succession of beds laid down in the progress of the ages, which constitutes *Stratigraphic geology.* The treatment of causes, agencies, and processes is the function of *Dynamic* or *Philosophic geology.*

Besides these there are special applications which give occasion for other terms, as *Economic geology,* which is concerned with the industrial applications of geologic knowledge; *Mining geology,* which is a sub-section of economic geology, relating to the application of geologic facts and principles to mining operations; *Atmospheric geology*, *Glacial geology*, and others that define themselves, and are for the greater part but limited aspects of the broad science.

Dominant processes.—Three sets of processes, now in operation on the surface of the lithosphere, have given rise to most of the details of its configuration, and even many of its larger features. These processes have been designated *diastrophism*, *vulcanism*, and *gradation*. *Diastrophism* includes all crustal movements, whether slow or rapid, gentle or violent, slight or extensive. Many parts of the land, especially along coasts, are known to be slowly sinking relative to the sea-level, while other parts are known to be rising. The fact that rocks originally formed beneath the sea now exist at great elevations, and the further fact that areas which were once land are now beneath the sea, are sufficient evidence that similar changes have taken place in the past. *Vulcanism* includes all processes connected with the extrusion of lava and other volcanic products, and with the rise of lava from lower to higher levels, even if not extruded. Vulcanism and diastrophism may be closely associated, for local movements at least are often associated with volcanic eruptions, and more considerable movements may be connected with the movements of subsurface lavas, even when the connection is not demonstrable. *Gradation* includes all those processes which tend to bring the surface of the lithosphere to a common level. Gradational processes belong to two categories—those which level down, *degradation*, and those which level up, *aggradation*. The transportation of material from the land, whether by rain, rivers, glaciers, waves, or winds, is *degradation* and the deposition of material, whether on the land or in the sea, is *aggradation*. Degradation affects primarily the protuberances of the lithosphere, while aggradation affects primarily its depressions.

ASTRONOMIC GEOLOGY.

The earth as a planet.—Though supremely important to us, the earth is but one of the minor planets attendant upon the sun, and is in no very special way distinguished as a planetary body. Of the eight planets, four, Jupiter, Saturn, Uranus, and Neptune, are much larger than the earth, while three, Mars, Venus, and Mercury, are smaller. There are a host of asteroids, but all together they do not equal the mass of the smallest planet. The average mass of the eight planets is more than fifty times that of the earth, while the largest, Jupiter, is more than three hundred times as massive as the earth. The earth's position in the group is in no sense distinguished. It is neither the outer nor the inner, nor even the middle planet. Even in the minor group to which it belongs, it is neither the outermost nor the innermost member, though in this group it is the largest. Its average distance from the sun is about 92.9 million miles, and this fixes its revolution at 365¼ days, for its period of revolution is directly dependent on its distance from the sun, and is necessarily longer than the revolutions of the inner planets and shorter than those of the outer planets. Its rotation in twenty-four hours is not far different from that of its neighbor Mars, but is much slower than the more distant and larger planets, Jupiter and Saturn, which rotate in about ten hours. Comparison cannot be made with the innermost and outermost planets, because their rotations are not yet satisfactorily determined. The plane of the earth's revolution lies near the common plane of the whole system, but this is not peculiar, as all of the planets revolve in nearly the same plane. Only a few of the small asteroids depart notably from this common plane. This has an important bearing on theories of the origin of the system, since this close coincidence of the planes of the orbits is not consistent with any haphazard aggregation of the material. Of similar importance is the fact that all of the planets revolve in the same direction and in ellipses that do not depart widely from circles. The eccentricity of the earth's orbit is only about $\frac{1}{60}$. This eccentricity varies somewhat, due to the disturbing influences of the other planets, and this variation has been regarded by some geologists as an influential cause of climatic changes, but its adequacy to produce great effects has been doubted by others. The inclination of the earth's axis, now about 23½°, holds an intermediate position, some of the planets having axes more inclined, as Saturn, $26\frac{5}{8}$°, and others less inclined, as Jupiter, 3°. The inclination of the axis is subject to trivial variations at present, and in the long periods of the past has possibly changed more notably. This possible change has also been thought to be a cause of climatic variation, but its efficiency has not been demonstrated.

Its satellite.—The earth is peculiar in having one unusually large satellite, which has a mass $\frac{1}{81}$ of its own. The great planets have several satellites whose combined mass exceeds that of the moon, and perhaps in some few cases the individual satellites may be larger than the moon, but they do not sustain so large a ratio to their planets, for Titan, probably the largest, is only $\frac{1}{4600}$ of the mass of Saturn. There is little doubt that the moon has played an important part in the history of the earth. It is the chief agency in developing oceanic tides, and it possibly also develops a body tide in the earth itself. These tides act as a brake on the rotation of the earth and tend to reduce its rate, and thereby to lengthen the day. While this may have been counteracted in some measure by the shrinkage of the earth, which tends to increase its rate of rotation, it has been held by eminent physicists and geologists that the rotation of the earth has been greatly lessened during its history, and that a long train of important consequences has resulted. If the contraction of the earth has been sufficient to offset this lessening, the tidal brake must be credited with the prevention of the excessive speed of rotation which would otherwise have been developed. The tides are efficient agencies in the shore wear of the oceans, and in the distribution of marine sediments, and these, it will be seen later, are important elements in the formation of strata.

Dependence on the sun.—By far the most important external relation of the earth, however, is its dependence on the sun. The earth is a mere satellite of the sun, less than $\frac{1}{300000}$ of its mass, and hence under its full gravitative control. The earth is dependent on the sun for nearly all its heat and light, and, through these, for nearly all of the activities that have given character to its history. It is too much to say that all activities on the surface of the earth are solely dependent on those of the sun, for a certain measure of heat and light and other energy is derived from other bodies, and a certain not inconsiderable source of energy is found in the interior of the earth itself; yet all of these are so far subordinate to that great flood of energy which comes from the sun that they are quite inconsequential. The history of the earth in the past has been intimately dependent upon that of the sun, and its future is locked up with the destiny of that great luminary. Geology in its broadest phases can therefore scarcely be separated from the study of the sun, but this falls within the function of the astronomer rather than the geologist.

Meteorites.—There are a multitude of small bodies passing through space in varying directions and with varying velocities and occasionally encountering the earth, to which they add their substance. Some of these meteorites revolve about the sun much as if they were minute planets, but some of them come from such directions and with such velocities as to show that they do not belong to the solar family. Some consist almost wholly of metal, chiefly iron alloyed with a small percent. of nickel (holosiderites); some consist of metal and rock intimately mixed (syssiderites and sporadosiderites); and some consist wholly of rock (asiderites). The rock is usually composed of the heavier basic minerals, though some meteorites consist largely of carbonaceous material. Besides meteorites, there is little doubt that wandering gaseous particles strike the earth, but this is beyond the reach of present demonstration. The amount of substance added to the earth by these meteorites and gases in recent times is relatively slight compared with the whole body of the earth. What contribution may have come to the earth in earlier times from such sources is a matter of hypothesis which will be discussed later.

GEOGNOSY.

The constitution of the earth.—Turning from its external relations to the earth itself, a natural threefold division is presented: (1) the atmosphere, (2) the hydrosphere, and (3) the lithosphere.

I. The Atmosphere.

The atmosphere is an intimate mixture of (1) all those substances that cannot take a liquid or solid state at the temperatures and pressures which prevail at the earth's surface, together with (2) such transient vapors as the various substances of the earth throw off. The first class form the permanent gases of the atmosphere, and consist of nitrogen about 79 parts, oxygen about 21 parts, carbon dioxide about .03 part, together with small quantities of argon, neon, xenon, krypton, helium, and other rare constituents. The second class are the transient and fluctuating constituents of the atmosphere, chief among which is aqueous vapor, which varies greatly in amount according to temperature, pressure, and other conditions. To this are to be added volcanic emanations and a great variety of volatile organic substances. Theoretically, every substance, however solid, discharges particles which may transiently become constituents of the atmosphere. Practically, only a few of these exist in such quantity as to be appreciable. Dust and other suspended matter are usually regarded as impurities rather than constituents of the atmosphere, but they play a not unimportant part by affecting its temperature and luminosity, and by facilitating the condensation of moisture.

Mass and extent.—The total mass of the atmosphere is estimated at five quadrillion tons, or $\frac{1}{1200000}$ of the mass of the earth. It is relatively dense at the surface of the earth and decreases in density outwards in a manner difficult of absolute determination, so that the actual height of the appreciable atmosphere is not positively known. The true conception of the atmosphere is perhaps that of a tenuous envelope exerting a pressure of about fifteen pounds per square inch at the sea-level, and thinning gradually upwards until it reaches a tenuity which is inappreciable, but perhaps not ceasing absolutely until the sphere of gravitative control of the earth is passed, about 620,000 miles from the lithosphere. In the lower portion, according to the kinetic theory of gases, the molecules fly to and fro, colliding with each other with almost inconceivable frequency, and with very short paths between successive collisions, but in the upper rare portion some of the molecules bound outwards, and do not strike other molecules, and hence pursue long elliptical paths until the gravity of the earth overcomes their momentum, when they return, perhaps to bound off again or to force other molecules to do so. This fountain-like nature of the outer part of the atmosphere makes any sharp definition of its limit impracticable. Some molecules are believed to be shot away at such speed that they do not return. Beyond about 620,000 miles from the surface of the lithosphere, the differential attraction of the sun is greater than that of the earth, and if the attraction of the earth does not turn the molecules back before reaching this distance, they are almost certain to be lost to the earth.

The measurement of heights by the aneroid barometer, which is much used in practical geology, is dependent on the lessening of pressure as the instrument is carried upward.

Geologic activity.—The atmosphere is the most mobile and active of the three great subdivisions of the earth, and when its indirect effects through the agency of water, as well as its direct effects, are considered, it is to be regarded as one of the most effective agencies of change. It acts chemically upon the rock substance of the earth, causing induration in some instances, but more often inducing disintegration and change of composition by means of which rock is reduced to soil, or soil-like material, and rendered susceptible of easy removal by winds and waters. When in motion the atmosphere acts mechanically on the surface of the earth, transporting dust and sand, and by the friction of these it abrades the surface. It is chiefly effective, however, in furnishing the conditions for water action. Partly by its mechanical aid, but chiefly by securing the right temperature, it is a necessary factor in the action of rains, streams, glaciers, and the various forms of moving water upon land. So also, on the ocean, wave action is essentially dependent on the winds. In the absence of atmospheric propulsion, wave action would be chiefly confined to the tides and to occasional earthquake impulses, and would lose nearly all its efficiency. Stream action and wave action, which are the most declared of the geological agencies, are therefore to be credited as much to the atmosphere as to the hydrosphere, since the action is a joint one to which both envelopes are essential.

A thermal blanket.—A function of the atmosphere of supreme importance is the thermal blanketing of the earth. In its absence the heat of the sun would reach the surface with full intensity, and would be radiated back from the surface almost as rapidly as received, and only a transient heating would result. During the night an intensity of cold would intervene scarcely less severe than the temperature of space. In penetrating the atmosphere certain portions of the radiant energy of the sun are absorbed. Of the remainder which reaches the surface of the earth, a part is transformed into vibrations of lower intensity, which are then more effectively retained by the atmosphere. The air thus distributes and equalizes the temperature. The two constituents of the atmosphere which are most efficient in this work are aqueous vapor and carbon dioxide, and the climate of the earth is believed to have been very greatly affected by the varying amounts of these constituents in the atmosphere, as well as by the total mass of the atmosphere.

The function of the atmosphere in sustaining life and promoting all that depends on life is too obvious to need comment.

The special geological action of the atmosphere will be discussed in the next chapter.

II. The Hydrosphere.

About 1300 quadrillion tons of water lie upon the surface of the solid earth. This equals about $\frac{1}{4640}$ part of the earth's mass. Were the surface of the solid earth perfectly spheroidal, this would constitute a universal ocean somewhat less than two miles deep. Owing to the inequalities of the rock surface, the water is chiefly gathered into a series of great basins or troughs occupying about three-fourths (72%)

of the earth's surface. These basins are all connected with each other and act as a unit, so that anything which changes the level of the water in one changes the level of all. This helps to make a common record of all great movements of the earth's body, for the level of the ocean determines where the detritus from the land shall lodge, and hence where the edge of the marine beds shall be formed. This will appear more clearly when the formation of marine strata is discussed.

Oceanic dimensions.—The surface area of the ocean is estimated by Murray at 143,259,300 square miles. Of this, somewhat more than 10,000,000 square miles lie on the continental shelf, i.e., lap up on the borders of the continental platforms. This shows that the great basins are somewhat more than full. If about 600 feet of the upper part of the ocean were removed, the true ocean basins would be just full, and the surfaces of the true continental platforms would be dry land. The area of the true oceanic basins is about 133,000,000 square miles, and that of the true continental platforms about 64,000,000 square miles. Under about 20% of the ocean area, the bottom sinks to depths between 6000 and 12,000 feet; under about 53% it sinks to depths between 12,000 and 18,000 feet; and under the remaining 4% it ranges from 18,000 feet down to about 30,000. The last includes those singular sunken areas known as "deeps," and sometimes called *anti-plateaus*, as they extend downward from the general ocean bottom much as the plateaus protrude upwards from the general land surface.

Besides the ocean, the hydrosphere includes all the water which constitutes the surface streams and lakes, together with that which permeates the pores and fissures of the outer part of the solid earth; but altogether these are small in amount compared with the great ocean mass.

Geologic activity.—Of all geological agencies water is the most obvious and apparently the greatest, though its efficiency is conditioned upon the presence of the atmosphere, upon the relief of the land, and upon the radiant energy of the sun. Through the agency of rainfall, of surface streams, of underground waters, and of wave action, the hydrosphere is constantly modifying the surface of the lithosphere, while at the same time it is bearing into the various basins the wash of the land and depositing it in stratified beds. It thereby becomes the great agency for the degradation of the land and the building up of the basin bottoms. It works upon the land partly by dissolving soluble portions of the rock substance, and partly by mechanical action. The solution of the soluble part usually loosens the insoluble, and renders it an easy prey of the surface waters. These transport the loosened material to the valleys and at length to the great basins, meanwhile rolling and grinding it and thus reducing it to rounder forms and a finer state, until at length it reaches the still waters or the low gradients of the basins and comes to rest. The hydrosphere is therefore both destructive and constructive in its action. As the beds of sediment which it lays down follow one another in orderly succession, each later one lying above each earlier one, they form a time record. And as relics of the life of each age become more or less imbedded in these sediments, they furnish the means of following the history of life from age to age. The historical record of geology is therefore very largely dependent upon the fact that the waters have thus buried in systematic order the successive life of the ages. Aside from this, the means of determining the order of events of the earth's history are limited and more or less uncertain.

The special processes of the hydrosphere in its various phases will be the subject of discussion hereafter (Chaps. III, IV, VI). Suffice it here to recognize its great function in the constant degradation of the land, and in the deposition of the derived material in orderly succession in the basins.

Chief horizons of activity.—The great horizons of geological activity are (1) the contact zone between the atmosphere and the hydrosphere, chiefly the surface of the ocean, (2) the contact zone between the hydrosphere and the lithosphere, chiefly the shore belts, and (3) the contact zone of the atmosphere and surface waters, with the face of the continents. It is in these three zones that the greatest external work is being done and has been done in all the known ages.

III. The Lithosphere.

The atmosphere and hydrosphere are rather envelopes or shells than true spheres, though in some degree both penetrate the lithosphere. The lithosphere, on the other hand, is a nearly perfect oblate spheroid with a polar diameter of 7899.7 miles, and an equatorial diameter of about 26.8 miles more. Its equatorial circumference is 24,902 miles. its meridional circumference 24,860 miles, its surface area 196,940,700 square miles, its volume 260,000,000,000 cubic miles, and its average specific gravity about 5.57. The oblateness of the spheroid is an accommodation to the rotation of the earth, the centrifugal force at the equator being sufficient to cause the specified amount of bulging there. Computations seem to indicate that the accommodation is very nearly what would take place if the earth were in a liquid condition, from which the inference has been drawn that it must have been in that condition when it assumed this form, and must have continued essentially liquid until it attained its present rate of rotation, since, if the earth once rotated at a much higher speed, the flattening at the poles and the bulging at the equator must have been correspondingly greater. It is thought by others, however, that the plasticity of the earth is such that it would at all times assume a close degree of approximation to the demands of rotation, even if the interior were in a solid condition. By still others it is thought that the contraction of the earth has tended to accelerate the rotation about as much as the tides have tended to retard it, and that it has undergone little change of form.

Irregularities.—It is only in a general view, however, that there is a close approximation to a perfect spheroidal surface. In detail there are very notable variations from it. Geodetic surveys seem to have shown that the equatorial diameters are not all equal, even when the measurements are reduced to sea-level, but research along this line has not reached a sufficient stage of completeness to permit satisfactory discussion. It is, however, highly probable that the ocean surface as well as the average land surface is warped out of the perfect spheroidal form to some notable degree. This is very likely due to inequalities in the density of the earth's interior. The fact that the larger portion of the water is gathered on one side of the globe, while the land chiefly protrudes on the opposite side, is very possibly due to unequal specific gravity in the interior of the earth.

The most obvious departure from a spheroidal form is found in the protrusion of the continents and in the sinking away of the earth surface under the oceans. As these inequalities present themselves to-day, they are known as continental platforms and ocean basins. These do not correspond accurately with the present land and water surfaces. About the continental lands there is a submerged border extending some distance out from the shore, and constituting a sea-shelf beyond which the surface descends rapidly to the great depths of the ocean. This slightly submerged portion, known as the continental shelf, belongs as properly to the continent as the adjacent low lands which are not submerged. The submergence of the edge of this shelf at present is usually about 100 fathoms, so that if the upper 600 feet

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of the ocean were removed, the outlines of the land would correspond quite closely with the border of the true continental platform.

It is customary to look upon the protrusions of the continents as the great features of the earth's surface, but in reality the oceanic depressions are the master phenomena. In breadth, depth, and capacity they much exceed the continental protrusions, and if the earth be regarded as a shrunken body, the settling of the ocean bottoms has doubtless constituted its greatest surface movement. From the estimates of Murray, Gilbert has derived the following tables, showing the relative areas of the lithosphere above, below, and between certain levels.[1]

From these estimates it appears that if the surface were graded to a common level by cutting away the continental platforms and dumping the matter in the abysmal basins, the average plane would lie somewhere near 9000 feet below the sea-level. The continental platform may be conceived as rising from this common plane rather than from the sea-level.

Contours.	Percent. of Surface above.	Percent. of Surface below.
Contour 24,000 feet above sea-level	0.004	99.996
" 18,000 " " "	0.09	99.91
" 12,000 " " "	0.7	99.3
" 6,000 " " "	2.3	97.7
ea-level	27.7	72.3
ontour 6,000 feet below sea-level	42.5	57.5
" 12,000 " " "	57.3	42.7
" 18,000 " " "	96.8	3.2
" 24,000 " " "	99.93	0.07

	Percent.
More than 6000 feet above sea-level	2.3
Between sea-level and 6000 feet above	25.5
Between sea-level and 6000 feet below	14.8
Between 6000 and 12,000 feet below sea-level	14.8
Between 12,000 and 18,000 feet below sea-level	39.4
Between 18,000 feet and 24,000 feet	3.1

Epicontinental seas.—Those shallow portions of the sea which lie on the continental shelf, and those portions which extend into the erior of the continent with like shallow depths, such as the Baltic

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[1] The Earth. Johnson's Encyclopædia. See also statement of Murray in Smith- an An. Rept., 1899, p. 312. Reprint from Brit. A. A. S., Dover meeting, 1899, and t. Geog. Mag., Vol. XV, 1899, p. 511.

Sea and the Hudson Bay, may be called epicontinental seas, for they really lie *upon* the continent, or at least upon the continental platform; while those other detached bodies of water which occupy deep depressions in the surface are to be regarded as true abysmal seas, as, for example, the Mediterranean and Caribbean seas and the Gulf of Mexico, whose bottoms are as profound as many parts of the true ocean basin itself.

Diversities of surface.—The bottoms of the oceanic basins are diversified by broad undulations which range through many thousands of feet, but they are not carved into the diversified forms that give variety to land surfaces. The ocean bottoms are also diversified by volcanic peaks, many of which rise to the surface and constitute isolated islands. Some of them have notable platforms at or near the surface, cut by the waves or built up by the accumulation of sediment and of coralline and other growths about them. Aside from these encircling platforms, the solid surface usually shelves rapidly down to abysmal depths, so that the islands constitute peaks whose heights and slopes would seem extraordinary if the ocean were removed.

The surface of the land is diversified in a similar way by broad undulations and volcanic peaks, and also by narrower wrinklings and foldings of the crust; but all of these irregularities have been carved into diversified and picturesque forms by subaërial erosion. In this respect the surface of the land differs radically from the bed of the sea. The agencies which have produced the continental platforms and abysmal basins, and the great undulations and foldings, as well as the volcanic extrusions that mark them, are yet subjects of debate. Here lie some of the most difficult problems of geology, but these cannot be stated with sufficient brevity to find a place here.

The surface mantle of the lithosphere.—The surface of the lithosphere is very generally mantled by a layer of loose material composed of soil, clay, sand, gravel, and broken rock. This loose material is sometimes known as *mantle rock*, and sometimes as *rock waste*. On the land, mantle rock is often composed of the disintegrated products of underlying rock formations. It represents the results of the recent action of the atmosphere, of water, of changes of temperature, and of other physical agencies acting on the outer part of the rock sphere. The surface of this mantle is being constantly removed by wind and water, but as constantly renewed by continued decomposition of the rock below. In some areas, especially in the northern part of North America and the northwestern part of Europe, the soil graduates down into an irregular sheet of mixed clay, sand, gravel, and bowlders, known as *drift*. From this and other evidence it is inferred that at a time not greatly antedating our own, ice, chiefly in the form of glaciers, spread extensively over the high latitudes of the northern hemisphere. In some parts of the earth the surface is still covered by fields of snow and ice, comparable to those which formed the drift. In still other places, especially along the flood plains of streams, the mantie rock consists of deposits made by streams which were unable to carry their loads of sediment to the sea.

The crust of the lithosphere—Much of the detritus washed down from the land finds its way to bodies of standing water, and beneath lakes and seas the mantle of loose material is made up largely of the gravel, sand, and mud derived from the land. Before deposition these materials are more or less assorted and arranged in layers by waves and currents. When consolidated they constitute rock. The weathering of the rocks of the land, the wearing away of the resulting detritus, and its deposition beneath standing water, are among the most important processes of geologic change.

On the land, the mantle of loose material is sometimes absent, and in such places the surface of solid rock of the crust appears. Bare surfaces of rock are most commonly seen where the topography is rough, especially on the slopes of steep-sided valleys and mountains, and on the slopes of cliffs which face seas or lakes. Solid rock, without covering of soil or loose material of any sort, is also frequently seen in the channels of streams, especially where there are falls or rapids.

We have but to note the effects of a vigorous shower on a steep slope, or of a swift stream on its channel, or of waves on the cliffs which face lakes and seas, to understand at least one of the reasons why loose materials are frequently absent from steep slopes The very general exposure of solid rock where conditions favor surface erosion suggests that rock is everywhere present beneath the soil or subsoil. Fortunately there is an easy way of testing the universality of the crust beneath the mantle. In all lands inhabited by civilized peoples there are numerous wells and other excavations ranging from a few feet to several hundred feet in depth, and occasional wells and mine-shafts reach depths of several thousand feet. Even in shallow excavations rock is often encountered, and in most regions excavations as much as two or three hundred feet deep usually reach rock, and no really deep boring has ever failed to find it. It may, therefore, be accepted as a fact that the upper surface of the solid rock is nowhere far below the surface.

Concerning the thickness of the crust, if there be any true crust at all, little is known by direct observation. The deepest valleys, such as the canyon of the Colorado, and the shafts and borings of the deepest mines and wells, give knowledge of nothing but rock. The deepest excavations extend rarely more than a mile below the surface. It is certain that rock of known kinds extends to far greater depths.

The interior.—Concerning the great interior of the earth, little is known except by inference. From the weight of the earth,[1] it is inferred that its interior is much more dense than its surface. From its behavior under the attraction of other bodies, it is believed to be at least as rigid as steel, and its interior cannot, therefore, be liquid, in the usual sense of that term. From the phenomena of volcanoes, and from observations on temperature in deep borings, it is inferred that its interior is very hot. Further inferences concerning its character are less simply stated, and will be referred to later.

The solid part of the earth is therefore composed of (1) a thin layer of unconsolidated or earthy material, a few feet to a few hundred feet in thickness, covering (2) a layer or zone many thousands of feet, and probably many miles, thick, composed of solid rock comparable to that exposed at numerous points on the surface, and (3) a central mass, to which the preceding layers are but a shell, composed of hot, dense, and rigid rock, the real nature of which is not known by observation.

Varieties of rock in crust.—If the mantle of soil, subsoil, and glacial rubbish were stripped from the land, the surface beneath would be found to be made up of a great variety of rocks, all of which may be grouped into two great classes. About four-fifths of the land surface would be of rock arranged in layers, and the other fifth would be of crystalline rock, generally without distinct stratification, and often bearing evidence of the effects of high temperature.

Stratified rocks.—The composition of most stratified rocks corresponds somewhat closely with the composition of sediments now being carried from the land and being deposited in the sea. Their arrangement in layers is the same, and the markings on the surfaces of the layers, such as ripple-marks, rill-marks, wave-marks, etc., are identical. Furthermore, the stratified rocks of the land, like the recent sediments of the sea, frequently contain the shells and skeletons of animals, and sometimes the impressions of plants. Most of the relics of life found in the stratified rocks belonged to animals or plants which lived in salt water. Because of their structure, their composition, their distinctive markings, and the remains of life which they contain, it is confidently inferred that most of the stratified rocks which lie beneath the mantle rock of the land were originally laid down in beds beneath the sea, and that the familiar processes of the present time furnish the key to their history.

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[1] Its specific gravity as a whole is about 5.57, and the specific gravity of its outer portion is about 2.7.

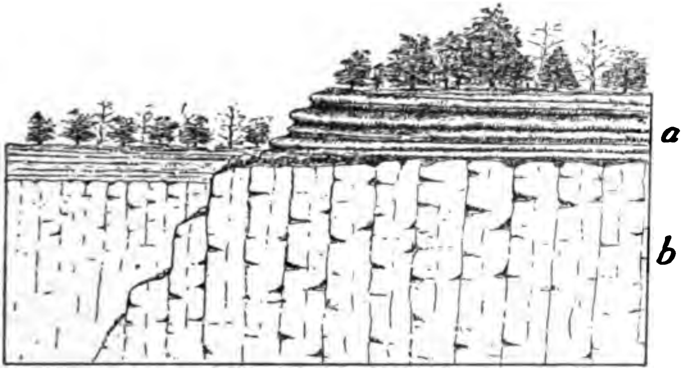

FIG. 1.—Beds of (Cambrian) sandstone, *a*, are conformable with one another, but unconformable on beds of (Huronian) quartzite, *b*. Near Ableman, Wis.

Conformability.—When the stratified rocks exposed by the removal of the mantle rock are examined, the successive beds are sometimes found to lie on one another in regular succession, showing that they were laid down one after another, without change in the attitude of the surface on which they were deposited. Such rocks are *conformable* (the beds of series *a*, Fig. 1). In other cases it would be seen that certain beds overlie the worn surfaces of lower beds, the layers of which may have a different angle of inclination (series *a*, Fig. 1, is unconformable on series *b*). Such relations show that the lower series of beds was disturbed and eroded before the overlying beds were deposited on them. Such series of rocks are *unconformable.*

Relative ages.—The structure and relations of rocks lead to inferences as to their relative ages. In the case of stratified rocks it is obvious that overlying beds were deposited later than those below, and where there is unconformity it is evident that an interval of time elapsed between the deposition of the unconformable series. Another and in some respects more important means of telling their order of formation is found in the remains of life entrapped in the water-laid sediments. Whatever life existed in the waters in which the sediments were deposited was liable to burial, and if it was possessed of hard parts, such as bones, teeth, shells, hard integuments, etc., these parts, or at least their impressions, were likely to be preserved in the sediments. Even tracks and imprints of perishable parts are sometimes preserved. All these relics, which we call *fossils*, give indications of the kinds of life which existed when the beds were formed. The fossils of the youngest beds show that the life which existed when they were deposited was quite like that of the present time. The fossils of the next older and lower beds show greater departure from present types. This series of changes continues downward as lower and lower beds are studied, until beds at considerable depths contain no relics of existing species but, in lieu thereof, forms of more primitive types. Some of these earlier types are clearly the ancestors of more modern forms, while others seem to have no living descendants. Going still deeper, the fossils indicate life of more and more primitive types, until they depart very widely from the living forms, and seem to be but remotely ancestral. So the beds may be followed downward until the lowest, which contain distinct evidences of life, are reached.

It should be understood that it is not possible to proceed directly downward through the whole succession of bedded rocks, but that the edges of the various beds may be found here and there where they have been brought to the surface by warpings or tiltings, or exposed by the wearing away of the beds which once overlay them. The full series of strata is made out only by putting together the data gathered throughout all lands, and even when this is done an absolutely complete series cannot yet be made out or, at least, has not been.

The crystalline rocks.—The crystalline rocks which would appear if the mantle rock were removed are of two types, *igneous* and *metamorphic*. Igneous rocks may be loosely defined as hardened lavas. Metamorphic rocks are those which are greatly changed from their original condition. Either stratified or igneous rocks may become metamorphic.

Igneous rocks sustain various relations to the stratified rocks, as illustrated by Fig. 2. From these relations it is possible to tell something of the order of their formation. Where the stratified rocks are broken through by lavas, it is obvious that the stratified rocks were formed first, and the lavas intruded later. Lava sheets intruded between beds of stratified rock can be told from those which flowed out on the surface and were subsequently buried, for in the former case the sedimentary rocks, both above and below the igneous rock, were affected by the heat, while in the latter case only those below were so affected.

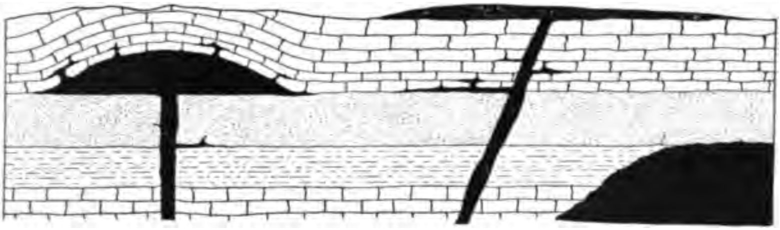

FIG. 2.—Diagrammatic representation of the relations of igneous rock to stratified rock. The igneous rocks, represented in black, have been forced up from beneath.

More commonly than otherwise the metamorphic rocks (Fig. 3) lie beneath the sedimentary beds and are often broken through by the igneous rocks. From their position in many places their great age may be inferred, but locally, especially where dynamic action has been severe, relatively young rocks are metamorphic.

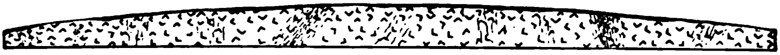

FIG. 3. The figure represents a section of the earth about 1000 miles long. The unequally thick black line at the top represents on something like its proper scale the depth of the stratified rock. The area below represents crystalline rock, largely metamorphic.

Four great sedimentary eras.—The water-laid series represents four great eras in the history of the earth, as shown by the relics of life imbedded in them. Beginning with the latest, these are the *Cenozoic* (recent life), during which the life took on its modern aspect; the *Mesozoic* (middle life), during which the life bore a mediæval aspect; the *Paleozoic* (ancient life), during which the life belonged to older types; and the *Proterozoic* (earlier life), during which it is inferred that much life prevailed, though its record is very imperfect. It may safely be assumed to have been more primitive than that of the Paleozoic, as it was earlier. Each of these great divisions embraced several lesser periods or epochs, and these again are subdivided more and more closely according to the degrees of refinement to which studies are carried. The chief of these subdivisions are given in the table on page 19, and others will come under consideration in the historical chapters.

In these four great series of sedimentary rocks there are, here and there, intrusions of igneous rocks, and in some places the sedimentary beds have been metamorphosed into crystalline rocks by heat and pressure. This is particularly true in the lowest of these series, the Proterozoic, where a large part of the sediment is metamorphosed, and where there is much igneous rock, but it is still clear that the main portion of this series was originally water-laid sediment, and so it belongs to the sedimentary series rather than the Archean, in which the sediments are the minor rather than the main factor. It has, however, usually been classed with the Archean, and it is certainly not always easy to draw the dividing line. In a sense it may be regarded as a transition series.

The Archean complex.—Beneath the dominantly sedimentary but partly metamorphic and igneous series there is a very complex group of rocks largely of metamorphosed igneous origin, though containing some metamorphosed sediments. These extend downwards to unknown depths. While all the great formations are occasionally bent and broken, these lowest ones are almost everywhere warped, folded, and contorted, often in the most intricate way. They have been very generally mashed and sheared by enormous pressure, so that they have become foliated, and their original character is much masked. They therefore form a series of great obscurity and complexity. As they are at the bottom of the known series, they have been called the "Fundamental gneiss" and the "Basement complex," but as the part which we see is not the true base nor the true foundation, it is safer to call them simply the Archean (very ancient) complex. As life appears to have been present during a part at least of the period of its formation is referred to the Archeozoic era.

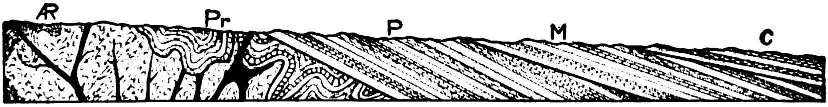

FIG. 4.—Diagram to illustrate the relations of the five great groups of formations. *Æ* = Archean, *Pr* = Proterozoic, *P* = Paleozoic, *M* = Mesozoic, *C* = Cenozoic.

Beyond and below this series, the structure of the earth is a matter of inference. Vast as are the preceding series, they together form relatively but a thin shell on the outer surface of the globe.

The foregoing series are diagrammatically expressed in Fig. 4, and natically presented to the eye in the following table.

GENERAL TABLE OF GEOLOGIC DIVISIONS.

CENOZOIC	Present.	
	Pleistocene.	
	Pliocene.	
	Miocene.	
	Oligocene.	
	Eocene	
	Transition (Arapahoe and Denver).	
MESOZOIC	Upper Cretaceous.	
	Lower Cretaceous (Comanche or Shastan).	
	Jurassic.	
	Triassic.	
PALEOZOIC	Permian.	
	Coal Measures or Pennsylvanian.	
	Subcarboniferous, or Mississippian.	
	Devonian.	
	Silurian.	
	Ordovician.	
	Cambrian.	
	Great interval.	
PROTEROZOIC	Keweenawan.	
	Interval.	
	Animikean or Penokean. (Upper Huronian of some authors).	
	Interval.	
	Huronian.	
	Great interval.	
ARCHEOZOIC	Archean Complex.	Great Granitoid Series. (Intrusive in the main, Laurentian.)
		Great Schist Series. (Mona, Kitchi, Lower Keewatin, Coutchiching, Lower Huronian of some authors.)

The purpose of this general survey is to bring the salient features of the earth's structure into view preparatory to entering in more detail into the study of particular processes and special formations and to lay a foundation for the fuller apprehension of the successive stages of the history of the earth, which constitutes the chief purpose of geological study. It is now advisable to turn to the detailed consideration of individual processes and specific structures. The complexity of the actions involved in the history of the earth is so great that such separate consideration at the outset is helpful.

CHAPTER II.

THE ATMOSPHERE AS A GEOLOGICAL AGENT.

WHILE it is convenient to regard the lithosphere as the earth proper, and the atmosphere as its envelope, the latter is as truly a part of the planet as the former, and its activities and its history are as truly subjects of geological study as the formation of the rocks. This view is in no way vitiated by the fact that the special study of the atmosphere is set apart under the name *Meteorology*, for in the same way the special study of rocks is set apart under the name *Petrology*, that of ancient life under the name *Paleontology*, and that of other phases of the subject under other names. The atmosphere is one of the three great formations of the earth, and as a geological factor takes its place beside the hydrosphere and the lithosphere. It has played a part in the history of the earth comparable to that of the water, though its mass is less and its record more elusive. Unsubstantial as the atmosphere seems when contrasted with the liquid and solid portions of the earth, its extreme mobility and its chemical activity compensate for its lightness and tenuity, and give it a function of the first order of importance.

The atmosphere plays a direct part as (1) a mechanical and (2) a chemical agent, and at the same time serves an indirect function in furnishing favorable conditions under which (3) solar radiation produces temperature effects, and (4) evaporation gives origin to precipitation and stream effects, and furnishes the necessary conditions for land plants and animals, and the important influences that spring from them.

This chapter is devoted to the work of the atmosphere in these and some less notable phases. The consideration of the origin and history of the atmosphere will receive attention later.

A. THE ATMOSPHERE AS A DIRECT AGENCY.

I. Mechanical Work.

The mechanical work of the atmosphere is accomplished chiefly through its movement. A feeble breeze is competent to move particles of dust, and winds of moderate velocity to shift sand. Exceptionally strong winds sometimes move small pebbles, but winds of sufficient force to move larger pieces of rock are rare. It follows, therefore, that the impact of the wind has little direct effect except on surfaces covered with dust and dry sand.

The transportation of material by the wind is limited by the size of the particles to which it has access. Dust particles expose more surface to the wind relative to their mass than sand grains. Winds which are unable to carry sand may still carry dust, and winds which are able to shift sand no more than a trivial distance may blow dust great distances.

The common conception of wind as a horizontal movement of some part of the atmosphere is not altogether accurate. Every obstacle against which wind blows causes deflections of its currents, and some of these deflections are upward. Furthermore, there are exceptional winds, in which the vertical element predominates. Particles of dust are often involved in these upward currents, and by them carried to great heights, and in the upper air are transported great distances.

Transportation and deposition of dust.[1]—The universality of the transportation of dust by the wind is well known. No house, no room, and scarcely a drawer can be so tightly closed but that dust enters it, and the movements of dust in the open must be much more considerable. The visible dustiness of the atmosphere in dry regions during wind-storms is adequate and familiar proof of the efficiency of the wind as a transporter of dust.

Under special circumstances, opportunity is afforded for rough determinations of the distance and height to which wind-blown dust is transported. Snow taken from snow-fields in high mountain regions is found to contain a small amount of earthy matter. Its particles are often found to be in part volcanic, even when the place whence the snow was taken is scores or even hundreds of miles from the nearest volcano. There is probably no snow-field so high, or so far from volcanoes, but that volcanic dust reaches it. If this be true of all snow-fields, it is probably true of all land surfaces. In the great Krakatoa[2] eruptions of 1883 large quantities of volcanic ash (pulverized lava) were projected to great heights into the atmosphere. The coarser particles soon settled; but, caught by the currents of the upper atmosphere, many of the finer particles were transported incredible distances. Through all their long journey, the particles of dust were gradually settling from the atmosphere, but not until the dust had traveled repeatedly round the earth did its amount become so small as to cease to make its influence felt in the historic red sunsets which it occasioned.[1] Some of this dust completed the circuit of the earth in 15 days.

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[1] For an excellent study of the erosion, transportation, and sedimentation performed by the atmosphere, see Udden, Jour. of Geol., Vol. II, pp. 318–331. See also Pop. Sci. Mo., September, 1896.

he Eruption of Krakatoa. Committee of the Royal Society, 1888.

In various parts of Kansas and Nebraska[2] there are very considerable beds of volcanic dust, locally as much as 30 feet thick, which must have been transported from volcanic vents by the wind, though there are no known centers of volcanic action, past or present, within some hundreds of miles of some of the localities where the dust occurs. These beds of volcanic dust, so far from its source, may serve as an illustration of the importance of atmospheric movements as a geological force.

Volcanic dust is shot into the atmosphere rather than picked up by it. Dust picked up by the wind is perhaps transported not less widely than volcanic dust, but, after settling, its point of origin is less readily determined. It would perhaps be an exaggeration to say that every square mile of land surface contains particles of dust brought to it by the wind from every other square mile, but such a statement would probably involve much less exaggeration than might at first be supposed.

Examples of extensive deposits of dust other than volcanic are also known. In China there is an extensive earthy formation, *the loess*, sometimes reaching 1,000 feet in thickness, which von Richtofen believes to have been deposited by the wind.[3] This conclusion has, however, not passed unchallenged.[4] The loess of some other regions has been referred to the same origin, and some of it is quite certainly eolian.[5]

The transportation of dust is important wherever strong winds blow over dry surfaces, free or nearly free of vegetation, and composed of earthy matter. Its effects may be seen in such regions as the sage-brush plains of western North America. The roots of the sage-brush hold the soil immediately about them, but between the clumps of brush, where there is little other vegetation, the wind has often blown away the soil to such an extent that each clump of brush stands up several inches, or even a foot or two, above its surroundings (Fig. 5). Such mounds are often partly due to the lodgment of dust about the bushes.

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[1] A brief account of the influence of the dust on sunsets is found in Davis's Elementary Meteorology, pp. 85 and 119.

[2] Science, New Ser., Vol. IV, p. 816, 1896.

[3] Von Richtofen. "China."

[4] Sketcherley and Kingsmill. Quar. Jour. Geol. Soc., Vol. LI, 1895, pp. 238–254.

[5] Chamberlin. Jour. of Geol., Vol. V, p. 795.

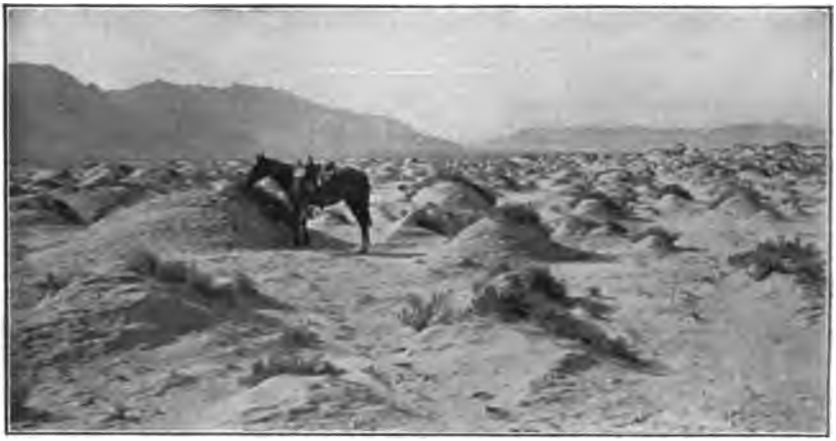

Fig. 5.—This figure shows the effect of sage-brush or other similar vegetation in holding sand or earth, or in causing its lodgment, in dry regions.

Where the earthy matter is moist, the cohesion of the particles is great, and the wind cannot pick them up. Furthermore, if the surface is generally moist, it is likely to be covered with vegetation which protects it against the wind. But even where vegetation is prevalent the wind finds many a vulnerable point. Thus on the edges of plains or plateaus facing abrupt valleys, the wind attacks the soil from the side, and in such situations all earthy matter may be stripped

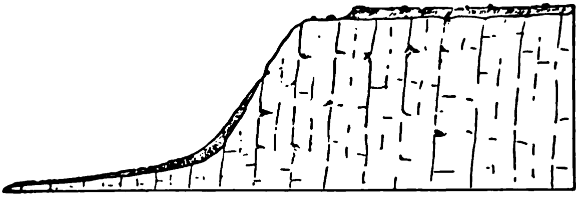

Fig. 6.—Diagram to illustrate the way in which the wind sometimes strips the soil from the edge of a bluff. This phenomenon is not rare in the basin of the Columbia River in Washington.

from the underlying rock for considerable distances from the edge of the cliff (Fig. 6). This may be seen at numerous points on the lava plateaus of Washington.

The presence of dust in the upper atmosphere during a rain-storm is sometimes the occasion of phenomena which are often misinterpreted. If there be abundant dust in the atmosphere through which rain-drops or snow-flakes fall, much of it is gathered up by them, and the water is thereby rendered turbid and the snow discolored. Here is to be found the explanation of "mud-rain," "blood-rain" (red dust), etc.

Since dust is carried to a considerable extent in the upper atmosphere, its movements and its deposition are little affected by obstacles on the surface of the land. A building or a hedge can only affect the lodgment of that part of the atmospheric dust which comes in contact with it or is swept into its lee. Since most obstacles on the surface of the solid part of the earth reach up but slight distances into the atmosphere, the dust of the greater part of the air settles without especial reference to them, and is spread more or less uniformly over the surface on which it falls.

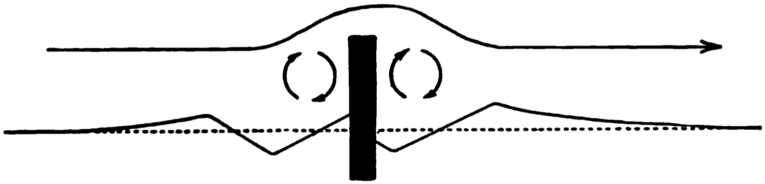

FIG. 7.—Diagram to illustrate the effect of an obstacle on the transportation and deposition of sand. The direction of the wind is indicated by the upper arrow. The lower arrows represent the direction of eddies in the air occasioned by the obstacle. If the surface in which the obstacle was set was originally flat (dotted line), the sand would tend to be piled up on either side at a little distance from the obstacle, but more to leeward. At the same time, depressions would be hollowed out near the obstacle itself (see full line). (After Cornish.)

Much of the dust transported by the wind is carried out over seas or lakes and falls into them. By this means, sedimentation is doubtless going on at the bottom of the whole ocean, and at the bottoms of all lakes. While means of determining the amount of dust blown into the sea are not at hand, it is safe to say that, were such determinations possible, the result, if stated in terms of weight, would be surprising.

Transportation and deposition of sand.—In its transportation by the wind, sand is not commonly lifted far above the surface of the land, and its movement is therefore more generally interfered with by surface obstacles than is the movement of dust. A shrub, a tree, a fence, a building, or even a stone may occasion the lodgment of sand in considerable quantity, though it has little effect on the lodgment of dust. The effect of obstacles is illustrated by Fig. 7 (see also Fig. 5). If the obstacle which occasions the lodgment of sand presents a surface which the wind cannot penetrate, such as a wall, sand is dropped abundantly on its windward side as well as on the leeward; but if it be penetrable, like an open fence, the lodgment takes place chiefly on its leeward side. In cultivated regions cases are known where, in a few weeks of dry weather, sand has been drifted into lanes in the lee of hedges to the depth of two or three feet, making them nearly impassable to vehicles.

Formation of dunes.—In contrast with dust deposited from the atmosphere, wind-blown sand is commonly aggregated into mounds and ridges in the process of lodgment. These mounds and ridges are dunes. Once a dune is started, it occasions the further lodgment of sand, and is a cause of its own growth. Dunes sometimes reach heights of 200 or 300 feet, but they are much more commonly no more than 10 or 20 feet in height. On plane surfaces, there is a limit in height above which they do not rise, though the limit is different under different conditions. The velocity of wind at the bottom of the air is not so great as that higher up, and as a dune is built up, a level is presently reached where the stronger upper winds sweep away as much sand as is brought to the top. The very even crests of many dune ridges are probably to be accounted for in this way. Wind-blown or eolian sand, not piled up in heaps or ridges, is somewhat widespread, but does not constitute dunes.

Shapes of dunes.[1]—Dunes may assume the form of ridges or of hillocks. The ridges may be transverse to the direction of the prevailing wind or parallel with it. Where dunes assume the form of hillocks rather than ridges, a group of them may be elongate in a direction parallel to the dominant wind, or at right angles thereto. The shape assumed by a dune or a group of dunes depends on the abundance of the sand, the strength and direction of the wind, and the shape of the obstacle which occasions the lodgment.

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[1] A thoroughgoing study of the *Formation of Sand Dunes* (by V. Cornish) is to be found in the Geog. Jour., Vol. IX, 1897, pp. 278–309.

The incipient stages of dune formation are readily seen in many dry, sandy regions. The dune is likely to start in the lee of some obstacle, and to be elongate in the direction of the wind, especially if the wind be strong relative to the supply of sand. This shape is per-

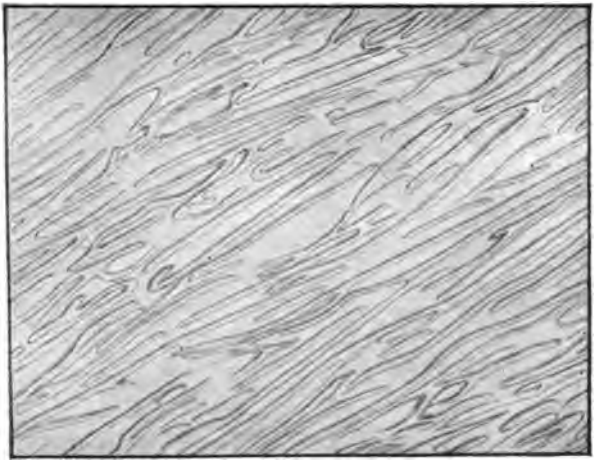

Fig. 8.—Dune ridges parallel to the direction of the wind. Southwest part of India. Scale about 3 miles to the inch. (Cornish.)

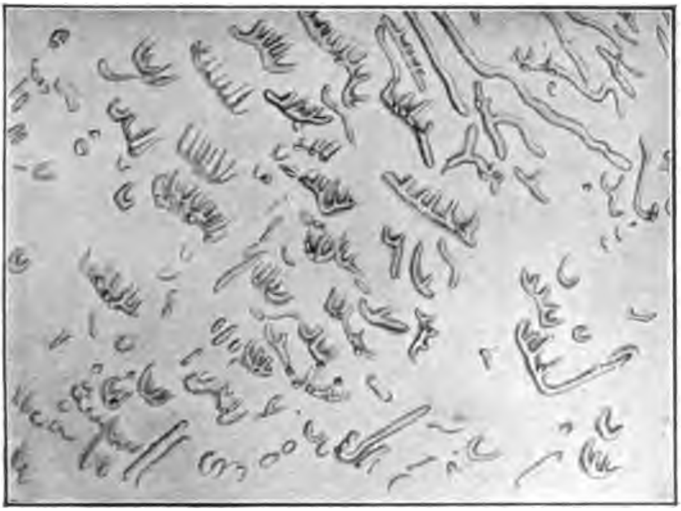

Fig. 9.—Dune ridges transverse to the direction of the wind. Scale about 3 miles to the inch. (Cornish.)

manently preserved if the proper relations between the supply of sand and strength and direction of wind are preserved. In the dune region of the Indian desert[1] the prevailing winds are alternately the southwest and northeast monsoons, the former being the stronger. The supply of sand comes from the southwest. Near the southwest coast the dune ridges are parallel to the direction of the wind (Fig. 8); in the interior, where the winds are less strong, the dunes are transverse to it (Fig. 9); while between the districts where these two types prevail intermediate forms occur. The transverse dune ridges (Fig. 9) are said to be the result of the lateral growth and erosion of longitudinal dunes.[2] In regions of changeable winds the shape of the dunes is subject to great variation. Dunes are sometimes crescentic, the convexity facing the wind (Fig. 10).

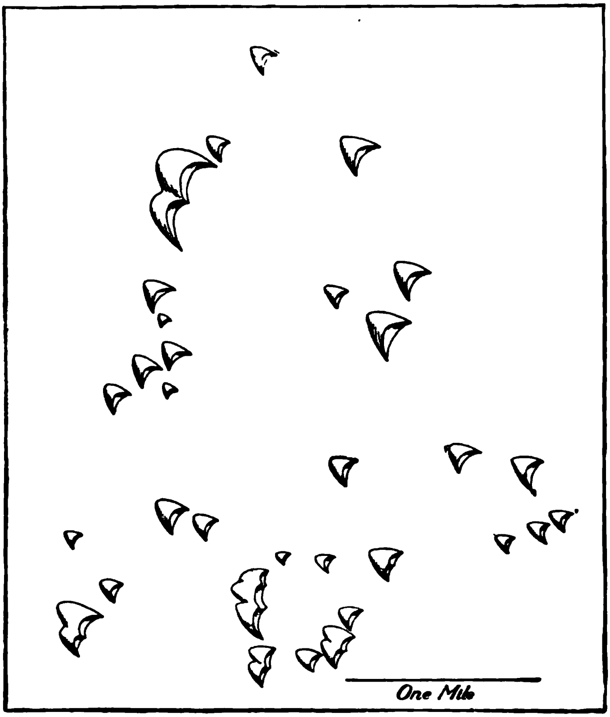

Fig. 10.—Crescentic dunes in ground-plan, the convexities facing the wind. (Bokhara.) (Walther.)

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[1] Blanford. Geology of India, 2d ed., p. 455 et seq.

[2] Cornish, loc. cit.

Along coasts, dune ridges are often transverse to the wind, and groups of dune hillocks are frequently elongate in the same direction. Here the source of supply of the sand is itself an elongate belt, often transverse to the dominant wind, and the resulting dunes often have great length transverse to the wind. Where the wind has strong mastery over the sand, the longitudinal tendency is seen, even along coasts.[1]

The shapes of dunes in section, like the shapes in ground-plan, depend on the relative strength and constancy of the winds and the supply of sand. With constant winds and abundant drifting sand, dunes are steep on the lee side (*bc*, Fig. 11), where the angle of slope is the angle of rest for the sand. It rarely exceeds 23° or 24°.[2] Under the same conditions the windward slope is relatively gentle (*ab*, Fig. 11). If the winds be variable so that the windward slope of one period becomes the leeward slope of another, and vice versa, this form is not preserved. Thus, by reversal of the wind, the section *abc*, Fig. 11, may be changed to *adc*. If the winds and the supply of sand be equal, on the average, from opposite directions, the slopes should, on the average, be equal, though perhaps unequal after any particular storm. The steep slopes of new-made dunes are lost after the sand has ceased to be blown. At some points where the winds erode (scour) more than they deposit, new profiles are developed (Figs. 12 and 13). The erosion profiles may be very irregular if the dunes are partially covered with vegetation. The effect of vegetation in restraining wind erosion is shown in Fig. 14, where plants have preserved a remnant of a dune.

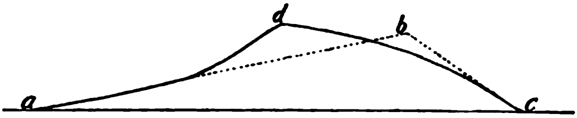

FIG. 11.—Section of a dune showing, by the dotted line, the steep leeward (*bc*) and gentler windward (*ab*) slope. By reversal of the wind the cross-section may be altered to the form shown by the line *adc*. (Cornish.)

FIG. 12.—Cross-section of a dune showing the profiles developed by scour of the wind on both flanks. (Cornish.)

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[1] Cornish, loc. cit., p. 294.

[2] Diller states (17th Ann. Rept., U. S. Geol. Surv., Pt. I, p. 450) that on the coast of Oregon the slope of dunes is sometimes 40°.

The topographic map.—Since dunes as well as other topographic features are conveniently represented on contour maps, and since such maps will be used frequently in the following pages, a general explanation of them is here introduced.

"The features represented on the topographic map are of three distinct kinds: (1) inequalities of surface, called *relief*, as plains, plateaus, valleys, hills, and mountains; (2) distribution of water, called *drainage*, as streams, lakes, and swamps; (3) the works of man, called *culture*, as roads, railroads, boundaries, villages, and cities.

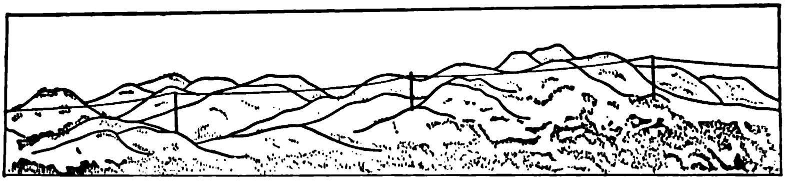

FIG. 13.—Diagram showing the outline of dunes in process of destruction. Seven Mile Beach, N. J. (N. J. Geol. Surv.)

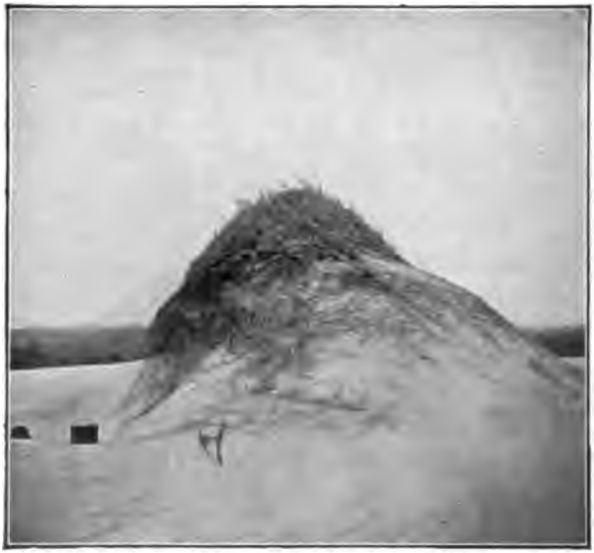

FIG. 14.—Illustrates the protective effect of vegetation against wind erosion. Dune Park, Ind. (Cowles)

Relief.—All elevations are measured from mean sea-level. The heights of many points are accurately determined, and those which are most important are given on the map in figures. It is desirable, however, to give the elevation of all parts of the area mapped, to delineate the horizontal outline, or contour, of all slopes, and to indicate their grade or degree of steepness. This is done by lines connecting points of equal elevation above mean sea-level, the lines being drawn at regular vertical intervals. These lines are called *contours*, and the uniform vertical space between each two contours is called the *contour interval*.

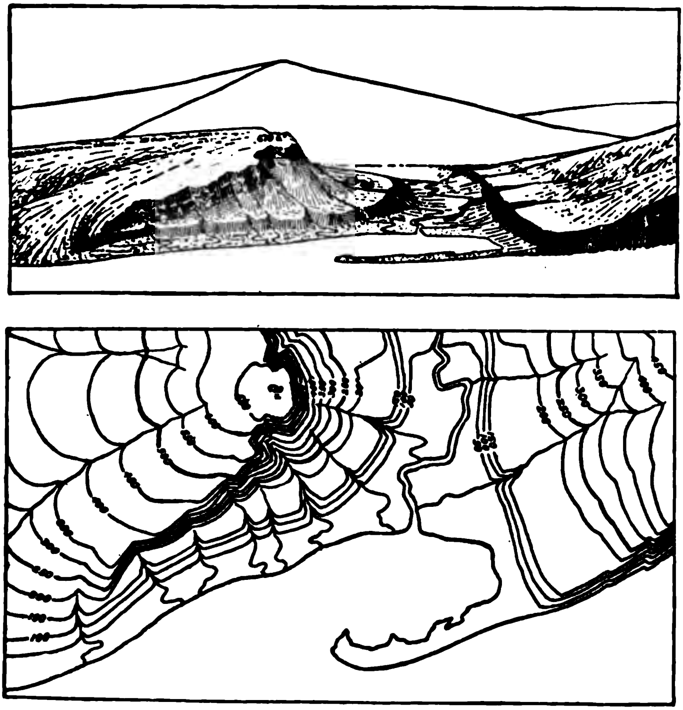

FIG. 15.—Sketch and map of the same area to illustrate the representation of topography by means of contour lines (U. S. Geol. Surv.)

On the maps of the United States Geological Survey the contours and elevations are printed in brown (see Plate II).

The manner in which contours express elevation, form, and grade is shown in the following sketch and corresponding contour map, Fig. 15.

The sketch represents a river valley between two hills. In the foreground is the sea, with a bay which is partly closed by a hooked sand bar. On each side of the valley is a terrace. From the terrace on the right a hill rises gradually, while from that on the left the ground ascends steeply in a precipice. Contrasted with this precipice is the gentle descent of the slope at the left. In the map each of these features is indicated, directly beneath its position in the sketch, by contours. The following explanation may make clearer the manner in which contours delineate elevation, form, and grade:

1. A contour indicates approximately a certain height above sea-level. In this illustration the contour interval is 50 feet; therefore the contours are drawn at 50, 100, 150, 200 feet, and so on, above sea-level. Along the contour at 250 feet lie all points of the surface 250 feet above sea; and similarly with any other contour. In the space between any two contours are found all elevations above the lower and below the higher contour. Thus the contour at 150 feet falls just below the edge of the terrace, while that at 200 feet lies above the terrace; therefore all points on the terrace are shown to be more than 150 but less than 200 feet above sea. The summit of the higher hill is stated to be 670 feet above sea; accordingly the contour at 650 feet surrounds it. In this illustration nearly all the contours are numbered. Where this is not possible, certain contours—say every fifth one—are accentuated and numbered; the heights of others may then be ascertained by counting up or down from a numbered contour.

2. Contours define the forms of slopes. Since contours are continuous horizontal lines conforming to the surface of the ground, they wind smoothly about smooth surfaces, recede into all reëntrant angles of ravines, and project in passing about prominences. The relations of contour curves and angles to forms of the landscape can be traced in the map and sketch.

3. Contours show the approximate grade of any slope. The vertical space between two contours is the same, whether they lie along a cliff or on a gentle slope; but to rise a given height on a gentle slope one must go farther than on a steep slope, and therefore contours are far apart on gentle slopes and near together on steep ones.

For a flat or gently undulating country a small contour interval is used; for a steep or mountainous country a large interval is necessary. The smallest interval used on the atlas sheets of the Geological Survey is 5 feet. This is used for regions like the Mississippi delta and the Dismal Swamp. In mapping great mountain masses, like those in Colorado, the interval may be 250 feet. For intermediate relief contour intervals of 10, 20, 25, 50, and 100 feet are used.

Drainage.—Watercourses are indicated by blue lines. If the streams flow the year round the line is drawn unbroken, but if the channel is dry a part of the year the line is broken or dotted. Where a stream sinks and reappears at the surface, the supposed underground course is shown by a broken blue line. Lakes, marshes, and other bodies of water are also shown in blue, by appropriate conventional signs.

Culture.—The works of man, such as road, railroads, and towns, together with boundaries of townships, counties and states, and artificial details, are printed in black."[1]

Topography of dune areas.—From what has been said, it is clear that the topography of dune regions may vary widely, but it is always distinctive. Where the dunes take the form of ridges (Fig. 1, Pl. II), the ridges are often of essentially uniform height and width for considerable distances. If there are parallel ridges, they are often separated by trough-like depressions. Where dunes assume the form of hillocks (Figs. 2 and 3, Pl. II), rather than ridges, the topography is even more distinctive. In some regions, depressions (basins) are associated with the dune hillocks. Occasionally they are hardly less notable than the dunes themselves. A somewhat similar association of hillocks and basins is locally developed by other means, but dunes are made up of sand and usually of sand only, while the composition of similarly shaped hillocks and depressions shaped by other agencies is notably different.

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[1] From folio preface, U. S. Geol. Surv.

In Fig. 1, Plate II (Five Mile Beach, 8 miles northeast of Cape May, N. J.), the contour interval is 10 feet. There is here but one contour line (the 10-foot contour), though this appears in several places. Since this line connects places 10 feet above sea-level, all places between it and the sea (or marsh) are less than 10 feet above the water, while all places within the lines have an elevation of more than 10 feet. None of them reaches an elevation of 20 feet, since a 20-foot contour does not appear. It will be seen that some of the elevations in Fig. 1 are elongate, while others have the form of mounds.

Fig. 2 (Pl. II) shows dune topography along the Arkansas River in Kansas (near Larned); Fig. 4, dune topography in Nebraska (Lat. 42°, Long. 103°), not in immediate association with a valley or shore; and Fig. 3 shows irregular ridge-like dunes at the head of Lake Michigan. In Fig. 2 the contour interval is 20 feet. All the small hillocks southeast of the river are dunes. Some of them are represented by one contour and some by two. The altitude of the region is considerable, the heavy contour representing an elevation of 2100 feet; but the dunes themselves are rarely more than 20 feet above their surroundings. In Fig. 4, where the contour interval is also 20 feet, there are, besides the numerous hillocks, several depressions (basins). These are represented by hachures inside the contour lines. In some cases there are intermittent lakes (blue) in the depressions. The heavy contour at Spring Lakes in this figure is the contour of 4300 feet. There are two depression contours (4280 and 4260) below it. The bottom of the depression is therefore lower than 4260, but not so low as 4240. In Fig. 3 the contour interval is 10 feet, and the dune ridges north of Miller are more than 50 feet high. The dune ridges here have helped to determine the position of this branch of the Calumet River, and have blocked its former outlet. The present drainage is to the westward.

Migration of dunes.—By the continual transfer of sand from its windward to its leeward side, a dune may be moved from one place to another, though continuing to be made up, in large part, of the same sand. In their migration dunes sometimes invade fertile lands, causing so great loss that means are devised for stopping them. The simplest method (employed in France and Holland) is to help vegetation to get a foothold in the sand. The effect of the vegetation is to pin the sand down. As a dune ridge along a coast travels inland, another may be formed behind it. Successions of dune ridges are thus sometimes formed.

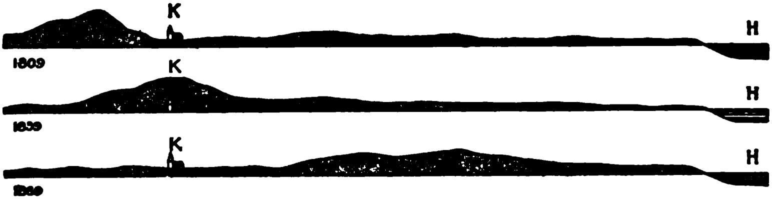

Fig. 16.—Diagram illustrating the migration of dunes on the Kurische Nehrung. (Credner.)

A remarkable instance of the migration of a sand dune is recorded on the Kurische Nehrung on the north coast of Germany. The Nehrung consists of a long narrow neck of land composed of sand, lying off the main coast. At the beginning of this century there was a notable dune ridge on one side. Since that time it has migrated a considerable distance, and in its migration it has been brought into the relationships illustrated in the accompanying diagrams (Fig. 16). In 1800

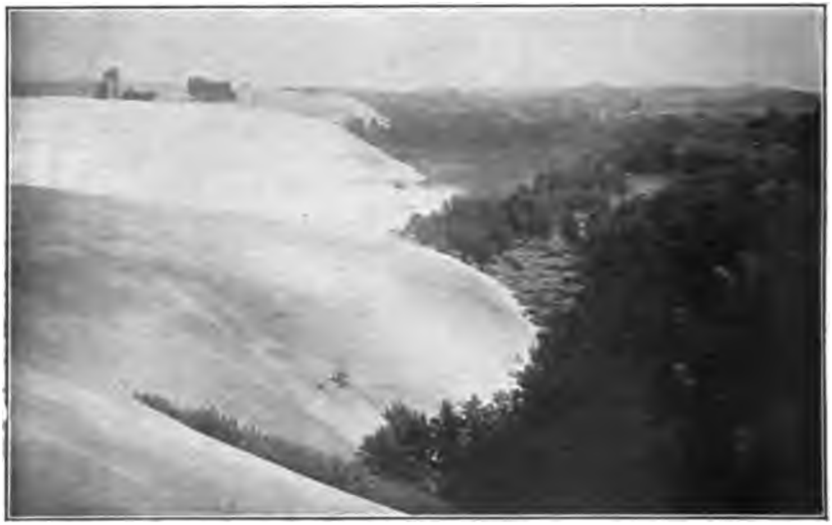

Fig. 17.—Migration of dunes into a timbered region. Dune Park, Ind. Head of Lake Michigan. (Meyers.)

the dune ridge was on one side of a church, which was then in use. In 1839 the ridge had been so far shifted to the leeward as to completely bury the church, and in 1869, its migration had progressed so far as to again discover the building.[1]

When dunes migrate into a timbered region they bury and kill the trees (Fig. 17). In one instance on the coast of Prussia a tall pine forest, covering hundreds of acres, was destroyed during the brief period between 1804 and 1827.[2] At some points in New Jersey orchards have been so far buried within the lifetime of their owners that only the tops of the highest trees are exposed. Trees and other objects once buried may be again discovered by farther migration of the sand (Figs. 18 and 19).[3]

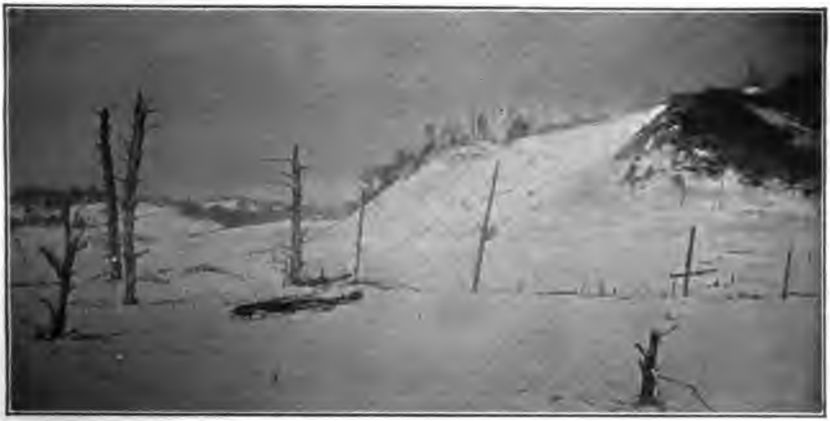

FIG. 18.—A resurrected forest. The dune sand after burying and killing the timber has been shifted beyond it. Dune Park, Ind. (Meyers.)

Eolian sand, not aggregated into distinct dunes, is often destructive. Even valleys and cities are sometimes buried by it. Drifting sands had so completely buried Nineveh two centuries after its destruction that its site was unknown.

Distribution of dunes.—Dunes are likely to be developed wherever dry sand is exposed to the wind. Their favorite situations are the dry and sandy shores of lakes and seas, sandy valleys, and arid sandy plains.

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[1] Credner. Elemente der Geologie, 6th ed., p. 271.

[2] Merrill. Rocks, Rock Weathering, and Soils, p. 295.

[3] Cowles. The Ecological Relations of the Vegetation of the Sand Dunes of Lake Michigan. Botanical Gazette, Vol. XXVII, 1899. An excellent study of the relations of sand dunes and vegetation.

Along coasts, dunes are likely to be extensively developed only where the prevailing winds are on shore. Thus about Lake Michigan, where the prevailing winds are from the west, dunes are abundant and large on the east shore, and but few and small on the west. In shallow water, shore currents and storm waves often build up a reef of sand a little above the normal level of the water. When the waves subside, the sand dries and the wind heaps it up into dunes. This sequence of events is in progress at many points on the Atlantic Coast. Sandy Hook, New Jersey, and the "beaches" farther south started as barrier ridges. When the waves had built them above normal water-level, the wind re-worked the sand, piling it up into mounds and hillocks (Fig. 1, Pl. II). Such dune belts a little off shore are sometimes turned to good account. They are usually separated from the mainland by a shallow lagoon. Where land is valuable, the lagoon is sometimes filled in, making new land, thus anticipating the result which nature would achieve more slowly. This has been done at some points on the western coast of Europe.

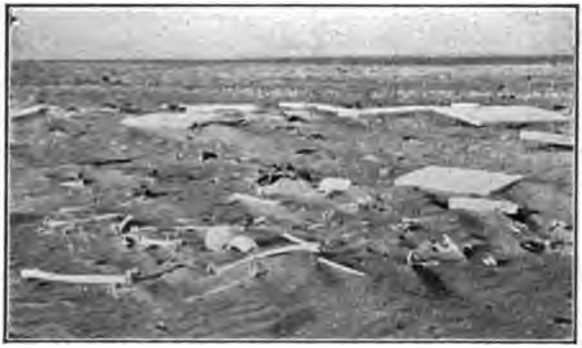

Fig. 19.—Migration of dune sand exposing bones in a cemetery. Hatteras Island, N. C. (Collier Cobb.)

Dunes are likely to occur along stream valleys (Fig. 2, Pl. II), if their bottoms or slopes are of sand, and not covered by vegetation. Dunes along valleys are usually on the side toward which the prevailing winds blow. Thus they are more common on the east side of the Mississippi than on the west. Dunes may be formed in the valley bottoms, but the sand is often blown up out of the valley and lodged on the bluffs above.

Apart from these special classes of situations, any sandy region the surface of which is dry is likely to have its surface material shifted by the wind and piled up into dune ridges or hillocks (Fig. 4, Pl. II). Dunes probably reach their greatest development in the Sahara, where some of them take the form of hillocks, and some the form of ridges. Travelers in that region report that dune ridges are sometimes encountered the faces of which are so high and steep as to be difficult of ascent, and that parties have been obliged to travel miles along their bases before finding a break where crossing was practicable.

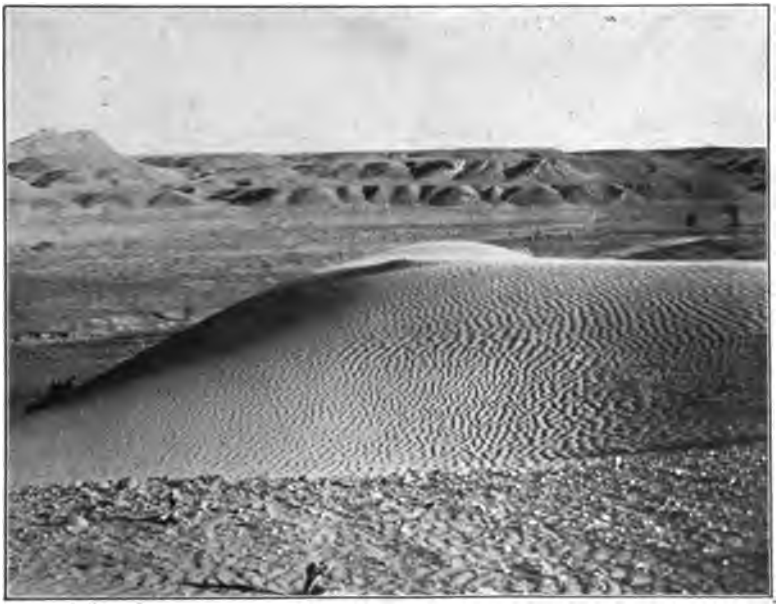

FIG. 20.—Wind-ripples. (Cross, U. S. Geol. Surv.)

Wind-ripples.—The surface of the dry sand over which the wind has blown for a few hours is likely to be marked with ripples (Fig. 20) similar to those made on a sandy bottom beneath shallow water, under the influence of waves. Like ripple marks made by the water, wind-ripples have one side (the lee) steeper than the other. While the ripples are, as a rule, but a fraction of an inch high, they throw much light on the origin of the great dune ridges. If the ripples be watched closely during the progress of a wind-storm, they are found gradually to shift their position. Sand is blown up the gentler windward slope to the crest of the ridge and falls down on the other side. The moment it falls below the crest of the ridge to leeward, it is protected against the wind, and is likely to lodge. Wear on the windward side is about equal to deposition on the leeward, and the result is the orderly progression of the ripples in the direction in which the wind is blowing, just as in the case of dune ridges.

Abrasion by the wind.—While the effect of the wind on sandy and dusty surfaces may be considerable, its effect on solid rock is relatively slight and accomplished, not by its own impact, but by that of the material it carries. The effect of blown sand on rock surfaces over and against which it is driven is perhaps best understood by recalling the effects of artificial sand-blasts, by means of which glass is etched. In a region where sand is blowing, exposed surfaces of rock suffer from a multitude of blows struck by the sand grains in transit. The result is that such rock surfaces are worn, and worn in a way peculiar to the agency accomplishing the work. If the rock be made up of laminæ which are of unequal hardness, the blown sand digs out the softer ones, leaving the harder projecting as ridges between them. Adjacent masses of harder and softer rock of whatever thickness are similarly affected. The sculpturing thus effected on projecting masses of rock is often picturesque and striking (Figs. 21 and 22), and is most common in arid regions. Details of wind-carving are shown in Fig. 23.

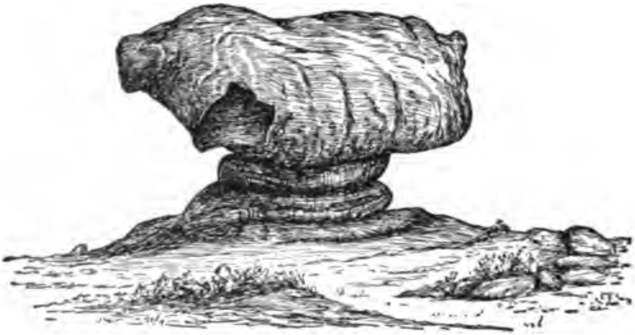

FIG. 21.—Wind-carved rock. (Green.)

PLATE II.

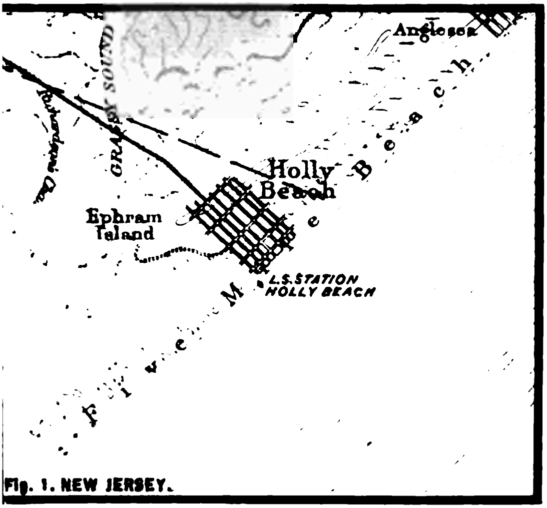

Fig. 1. NEW JERSEY.

U. S. Geol. Surv.

Scale, 1+ mile per inch.

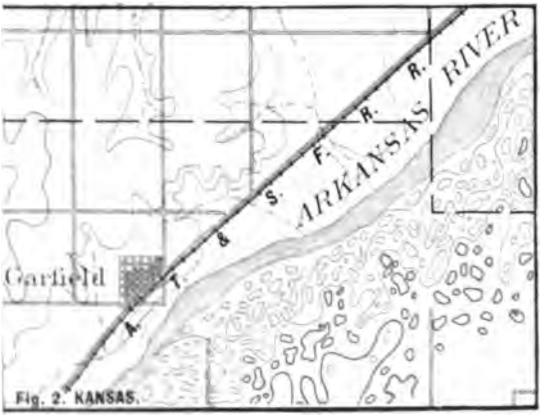

Fig. 2. KANSAS.

U. S. Geol. Surv.

Scale, 2+ miles per inch.

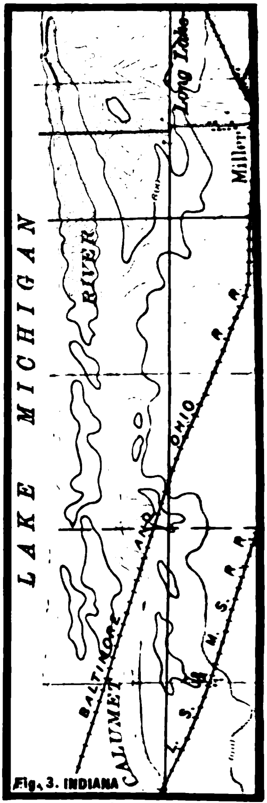

Fig. 3. INDIANA

U. S. Geol. Surv.

Scale, 1+ mile per inch.

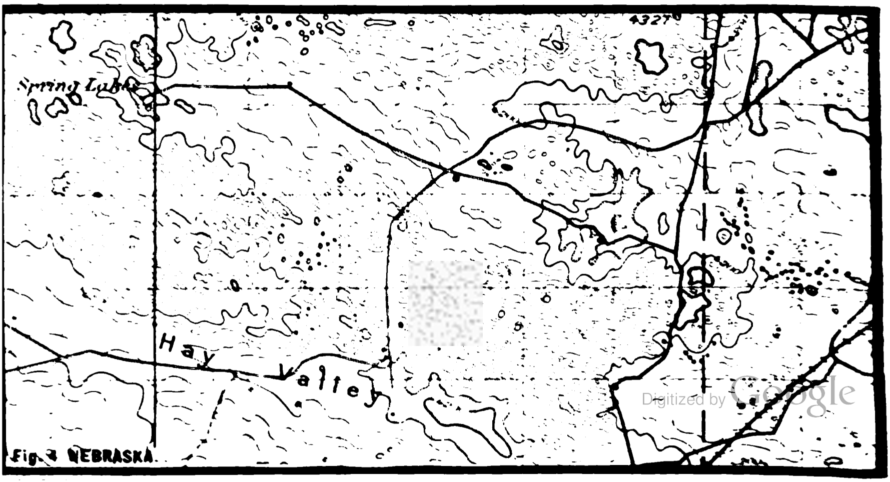

Fig. 4 NEBRASKA.

Sand drifted over loose stones lying on the surface often develops flat or flattish faces or facets on them. These facets are likely to be three in number, and the exposed portion of the stone is likely to develop a sort of pyramidal shape, the three flattish surfaces being mutually limited by tolerably well-defined lines (Fig. 24). Thus arise the three-faceted stones (*Dreikanter* of the Germans) commonly seen where sands have been long in movement.

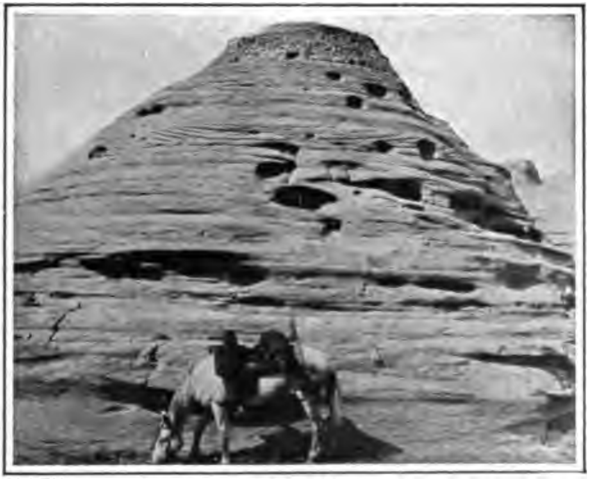

FIG. 22.—Wind-carved hillock of cross-bedded sandstone. Missouri River, Montana. (Calhoun, U. S. G. S.)

Not only does the drifting sand wear the surface over which it passes and against which it strikes, but the grains themselves are worn in the process. They are liable to be broken as they strike rock surfaces, and they are likely to strike one another in the atmosphere. In both cases they are subject to wear, and so to reduction to a finer and finer state.

The erosion accomplished by the wind is therefore of various sorts. The impact of the wind itself picks up the fine materials which are already loosened, thus wearing down the surface from which they are removed; the materials picked up wear the rock surfaces against which they are blown, and the transported materials themselves suffer reduction in transit.

Effects of wind on plants.—Another effect of wind work is seen in the uprooting of trees (Fig. 25). The uprooting disturbs the sur-

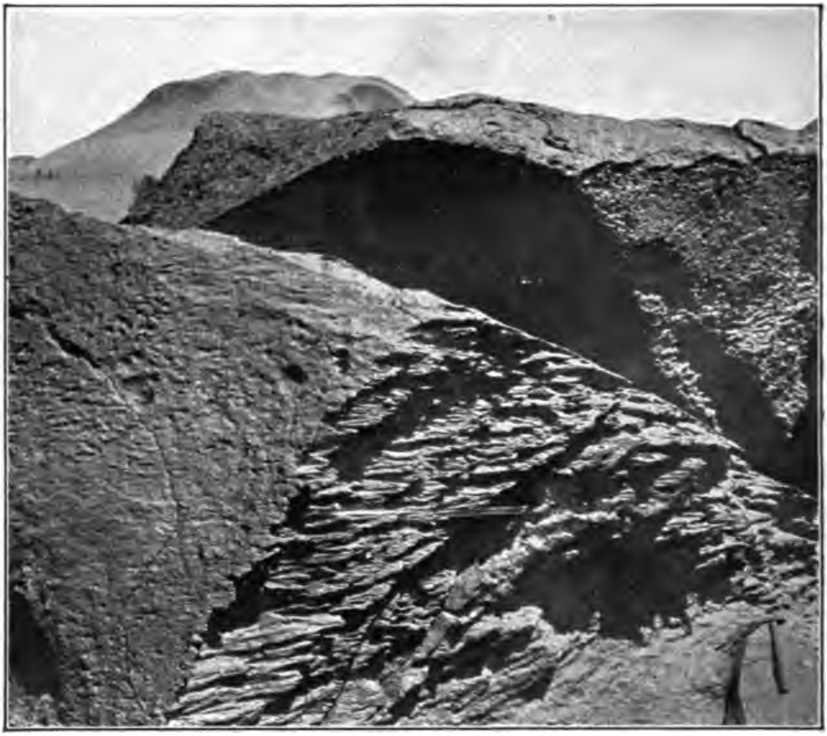

Fig. 23.—Figure showing details of wind-carving on rock surface (rhyolite). Mono Valley, California.

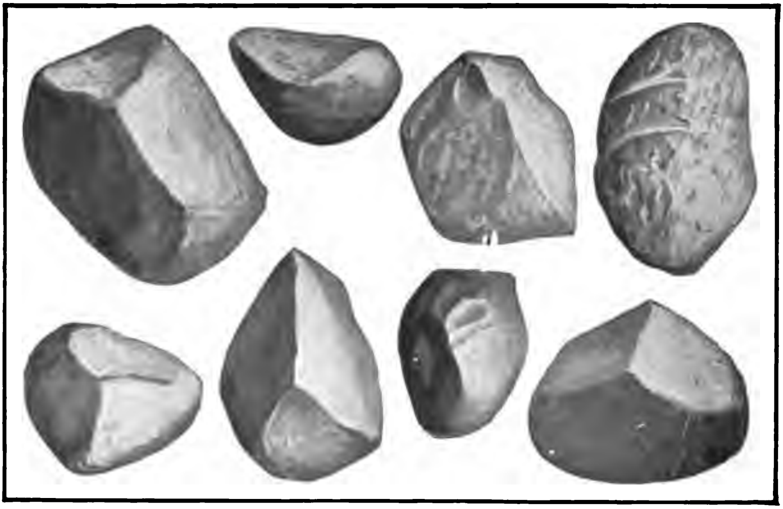

Fig. 24.—Wind-worn stones (*Dreikanter*).

face in such a way as to make loose earth more readily accessible to wind and water. The uprooting of trees on steep slopes often causes the descent of considerable quantities of loose rock and soil. Again, organisms of various sorts (certain types of seeds, germs, etc.), as well as dust and sand, are extensively transported by the wind. While this is important biologically its geological effects are remote.

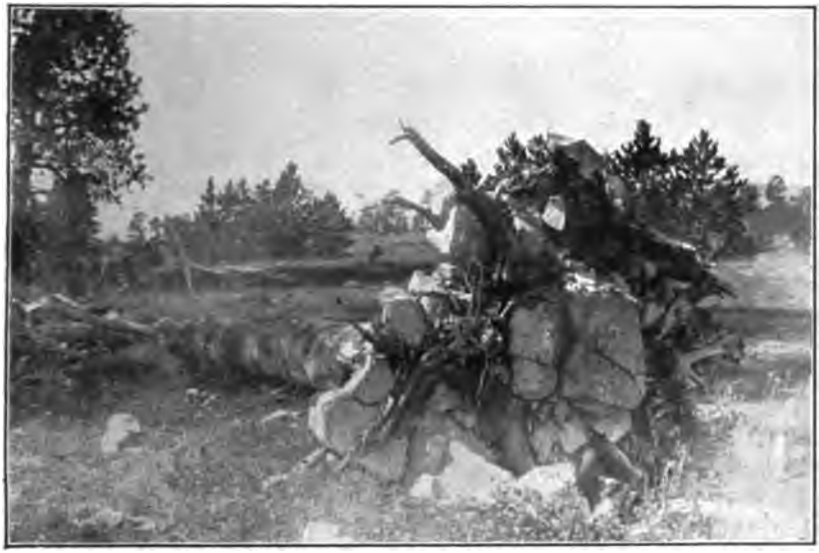

Fig. 25.—Shows the disturbance of surface earth and rocks by upturning of trees. (Darton, U. S. Geol. Surv.)

Indirect effects of the wind.—Other dynamic processes are called into being or stimulated by the atmosphere. Winds generate both waves and currents, and both are effective agents in geological work. The results of their activities are discussed elsewhere.

II. The Chemical Work of the Atmosphere.

The chemical work of the atmosphere (including solution and precipitation from solution) is principally accomplished in connection with water, a dry atmosphere having relatively little direct chemical effect on rock or soils.

Precipitation from solution.—The water in the soil is constantly evaporating. Such substances as it contains in solution are deposited where the water evaporates, and where evaporation is long continued without re-solution of the substances deposited, the surface becomes coated with an efflorescence of mineral matter. Conspicuous examples are found in the alkali plains of certain areas in the western part of the United States. Since the alkaline efflorescence is the result of evaporation it is connected with the atmosphere, but the material of the efflorescence was brought to its present position by water. The principle involved is illustrated by the white efflorescence which frequently appears on brick walls during the dry days which follow a drenching rain. The water penetrates the brick and mortar and dissolves something of their substance, and when it is evaporated from the surface the material in solution is left behind.

In arid regions the deposition of substances other than alkali is common. The percolating waters dissolve whatever is soluble, and when they evaporate their mineral content is left. The pebbles and stones of the arid plains have in many places become heavily coated with mineral matter deposited in this way, and not infrequently cemented into conglomerate. One of the commonest mineral substances found in such situations is lime carbonate. In some cases it was doubtless derived by solution from limestone beds beneath the surface, but this is not always the case. It often encrusts the bits of lava on lava plains where it can hardly have been derived from limestone. The faces of cliffs of granite or gneiss, hundreds and even thousands of feet above all other sorts of rock,[1] are sometimes spotted with patches of lime carbonate. In the first case the lime carbonate was derived by chemical change from the lava, and in the second, from the granite or gneiss (see *Carbonation* below), but its present position is the result of evaporation.

Oxidation.—In the presence of moisture the oxygen of the air enters into combination with various elements of the soil and rocks. This is *oxidation*. No other common mineral substance shows the results of oxidation so quickly and so distinctly as iron. The oxidized portion is loose and friable, and a mass of iron exposed to a moist atmosphere will ultimately crumble away. This change is comparable to other less obvious changes taking place in many minerals at and below the surface. Oxidation generally involves the disintegration of the rock concerned. Its effects in this direction will be referred to in other connections.

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[1] For example, in the Big Horn Mountains of Wyoming.

Carbonation.—The production of lime carbonate from rock containing calcium compounds, but not in the form of carbonates, is known as *carbonation*, and is one of the important chemical changes effected by the carbon dioxide of the atmosphere in coöperation with water. In the process of carbonation the original minerals of complex composition are decomposed and simpler ones usually formed. Volumetric changes are involved, which often lead to the disruption of the rock (see *Ground water*). Furthermore, carbonates are among the more soluble minerals, and their production therefore brings some of the rock materials into a soluble condition, and their extraction through solution tends still further to disintegrate the rock. The carbonation of crystalline rocks is therefore a disintegrating process, and will be considered further in its many concrete applications.

Other chemical changes.—A third chemical process which often accompanies oxidation and carbonation is *hydration*. This is effected by water rather than by air, and will be considered in that connection. In general it leads to the disintegration of the minerals and rocks affected. The chemical effects of nitric acid, etc., developed through the agency of atmospheric electricity, and the corresponding effects of the gases and vapors which issue from volcanoes, many of them chemically active, are to be mentioned in this connection.

Conditions favorable for chemical changes.—Conditions are not everywhere equally favorable for the chemical work of the atmosphere. In general, high temperatures facilitate chemical action, and, other things being equal, rocks are more readily decomposed by atmospheric action in warm than in cold regions. Chemical activity is probably greater where the climate is continuously warm than where there are great changes of temperature. Changes of temperature, on the other hand, tend to disrupt rock, and thus increase the amount of surface exposed to chemical change. Since nearly all the chemical changes worked by the atmosphere on the rocks are increased by the presence of moisture, the chemical activity of the atmosphere is greater in moist than in dry regions.

B. THE ATMOSPHERE AS A CONDITIONING AGENCY.

The most obvious mechanical work of the atmosphere is effected by the wind, but mechanical results of great importance, conditioned by the atmosphere, are also effected when the air is still.

I. Temperature Effects.

When the sun shines on bare rock its surface is heated and expanded, and the expanded particles crowd one another with great force. Since rock is a poor conductor of heat its surface is heated and expanded notably more than parts beneath the surface. It follows that strains are set up between the expanded outer portion and the cooler and less expanded parts within. In the cooling of the same rock mass it is the outermost portion which cools first and fastest, and, contracting as it cools, strains are again set up between the outer part, which is cooled more, and the inner part, which is cooled less. The result may be illustrated by the effect of cold water on hot glass, or of hot water on cold glass. In either case the fracture is the result of the sudden and considerable differential expansion or contraction. Since the heating and cooling of rock are much slower than the heating and cooling of glass under the conditions mentioned, the rupturing effects are less conspicuous, but none the less real. The actual effects of temperature changes are illustrated by familiar phenomena. The surface portions of bowlders exposed to the sun are frequently seen to be shelling off (Fig. 26). The loosened concentric shells may be a fraction of an inch, or sometimes even several inches in thickness. This process of *exfoliation* affects not only bowlders, but bare rock surfaces wherever exposed to the sun (Figs. 27, 28). It is often conspicuous on the faces of cliffs.

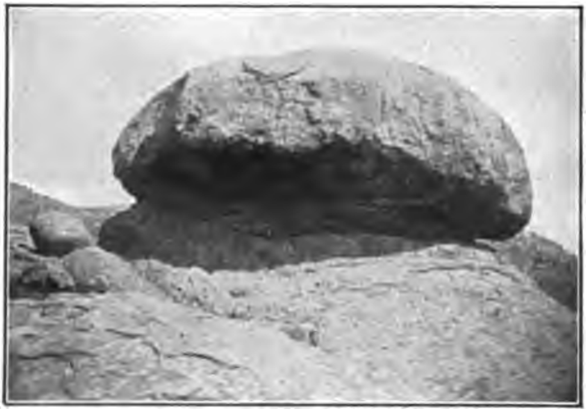

Fig. 26.—Exfoliation. A bowlder of weathering, the rock being granite. Wichita Mountains, Oklahoma.

Several conditions, some of which are connected with the atmosphere and some with the rock, determine the efficiency of this process. Since the breaking of the rock results from the expansion and contraction due to its changes of temperature, it follows that, other things being equal, the greater the change, the greater the breaking; but the suddenness of the temperature change is even more important than its amount. It follows that great daily, rather than great annual, changes of temperature[1] favor rock-breaking, though with changes of a given frequency their effectiveness is greater the greater their range. A partial exception to this generalization should be noted. If abundant moisture is present in the pores and cracks of the rock a change of temperature from 45° to 35° (Fahr.) might be far less effective in breaking the rock than a change from 35° to 25° in the same time, for in the latter case the sudden and very considerable expansion (about one-tenth) which water undergoes on freezing is brought into play. This may be called the *wedge-work of ice*. The daily range of temperature is influenced especially by latitude,

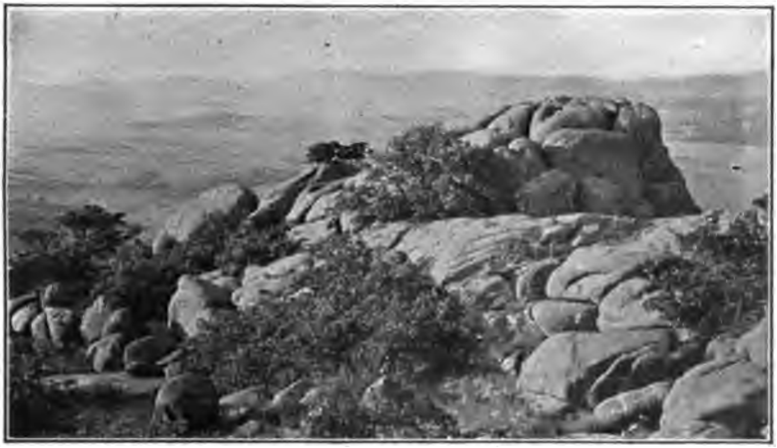

Fig. 27.—A weathered summit of granite in the Wichita Mountains, Oklahoma. (Willis, U. S. Geol. Surv.)

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[1] It should be noted that it is the change of temperature of the rock surface, not the change of temperature of the air above it, which is to be considered. Many data concerning temperature changes are to be found in Bartholomew's Atlas of Meteorology.

altitude, and humidity. Other things being equal, the greatest daily ranges of temperature occur in high-temperate latitudes, though to this general statement there are local exceptions, depending on other conditions. High altitudes favor great daily ranges of temperature, so far as the rock surface is concerned (see Figs. 29, 30), for though the rock becomes heated during the sunny day, the thinness and dryness of the atmosphere allow its heat to radiate rapidly at night. Here, too, the daily range of temperature is likely to bring the wedge-work of ice into play. Since the south side of a mountain (in the northern hemisphere) is heated more than the north, it is subject to the greater daily range of temperature, and the rock on this side suffers the greater disruption. Similarly, rock surfaces on which the sun shines daily are subject to greater disruption than those much shielded by clouds. Isolated peaks, because of their greater exposure, are subject to rather greater daily ranges of temperature than plateaus of the same elevation.

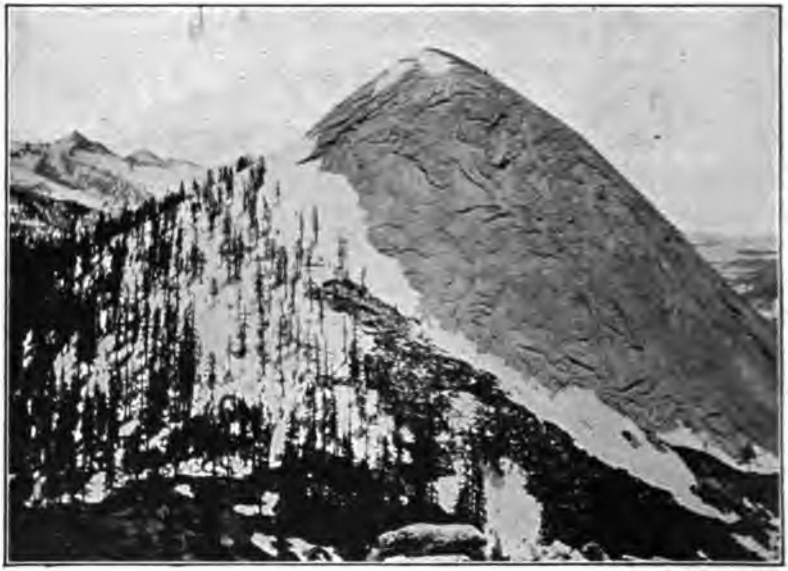

Fig. 28.—Exfoliation on a mountain slope. Mt. Starr-King (Cal.) from the north.

The daily range of temperature is also influenced by humidity. Because of the effect of water vapor in the atmosphere on insolation and radiation, a rock surface becomes hotter in the day and cooler at night beneath a dry atmosphere than beneath a moist one. Aridity therefore favors the disruption of rock by changing temperatures.

Turning from the conditions of the atmosphere which affect the disruption of rock to the conditions of the rock which influence the same process, several points are to be noted. In the first place, the disrupting effects of changes of temperature are slight or nil where the solid rock is protected by soil, clay, sand, gravel, snow, or other incoherent material. If the constituent parts of the loose material are coarse, like bowlders, their surfaces are affected like those of larger bodies of rock. The color of rock, its texture and its composition, also influence its range of daily temperature by influencing absorption and conduction. Dark-colored rocks absorb more heat than light-colored ones, and compact rocks are better conductors than porous ones. Great absorp-

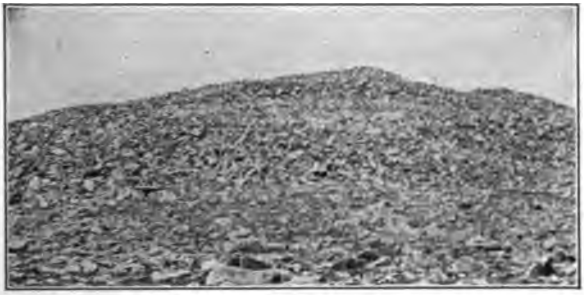

Fig. 29.—Top of Notch Peak, Bighorn Mountains, Wyo. Shows the thoroughly broken character of the rock on the summit, the absence of soil, vegetation, etc. (Kümmel.)

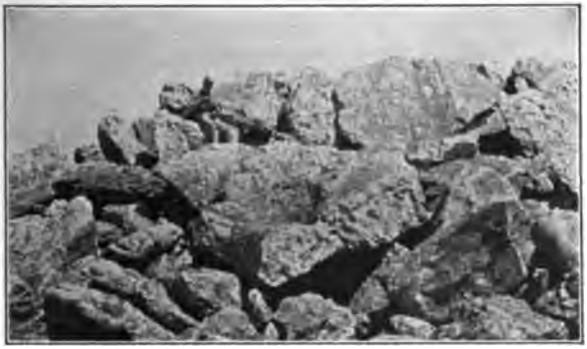

Fig. 30.—A detail from Fig. 29 showing the size of the rock blocks. (Kümmel.)

tion and slow conduction favor disruption. A given range of temperature is unequally effective on rocks of different mineral composition. In general crystalline rocks (igneous and metamorphic) are more subject to disruption by this means than sedimentary rocks, partly because they are more compact, but especially because they are made up of aggregates of crystals of different minerals which, under changes of temperature, expand and contract at different rates, while the common sedimentary rocks are made up largely of numerous particles of one mineral.

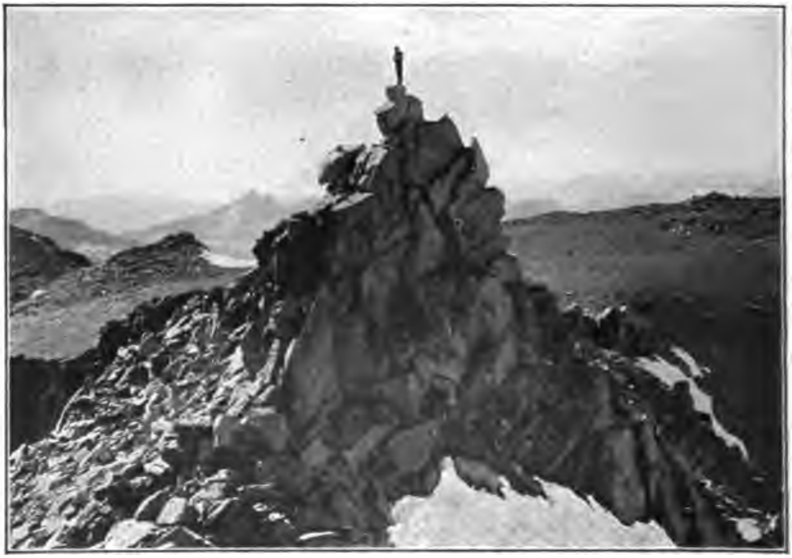

FIG. 31.—Peak north of Kearsarge Pass, the Sierras. Shows the way in which serrate peaks break up into angular blocks.

The freezing of water in the pores of rock is effective in disrupting them only when the pores are essentially full at the time of freezing. Otherwise there is room for the expansion attending the freezing. If the pores of the rock are large, the expansion on freezing may force out sufficient water to balance the increase of volume, even though the rock was completely saturated. If the pores be very small the water passes out less readily, and if the rock is saturated, freezing is more likely to be attended with disruption.[1]

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[1] Buckley. Wisconsin Survey, Bull. IV, 1899, pp. 81–3.

In view of these considerations the breaking of rock by changes of temperature should be greatest on the bare slopes of isolated elevations of crystalline rock, where the temperature conditions of temperate latitudes prevail, and where the atmosphere is relatively free from moisture. All these conditions are not often found in one place, but the disrupting effects of changing temperatures are best seen where several of them are associated (Figs. 29, 30, and 31).

The importance of this method of rock-breaking has rarely been appreciated except by those who have worked in high and dry regions. Climbers of high mountains know that almost every high peak is covered with broken rock to such an extent as to make its ascent dangerous to the uninitiated. High serrate peaks, especially of crystalline rock, are, as a rule, literally crumbling to pieces (Fig. 31). The piles of talus which lie at the bases of steep mountain slopes are often hundreds of feet in height, and their materials are often in large part the result of the process here under discussion. In mountain regions where atmospheric conditions favor sudden changes of temperature, the sharp reports of the disruption of rock masses are often heard. Masses of rock, scores and even hundreds of pounds in weight, are frequently thus detached and started on their downward course.[1] Small pieces of rock are of course much more commonly broken off than large ones. The disruption of rock by changes of temperature is not usually the result of a single change of temperature, but rather of many successive expansions and contractions.

The sharp needle-like peaks which mark the summits of most high mountain ranges (Fig. 32) are largely developed by the process here outlined. The altitude at which the serrate topography appears varies with the latitude, being, as a rule, higher in low latitudes and lower in high. But even in the same latitude it varies notably with the isolation of the mountains and with the aridity of the climate. Thus within the United States the sharply serrate summits appear in some places, as in Washington and Oregon, at altitudes of 6000 to 10,000 feet, while in the isolated Wichita range of Oklahoma, much farther south, but in a much drier climate, the same sort of topography is developed at altitudes of 2500 to 3000 feet.

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[1] Livingstone has reported that the temperature of rock surfaces in Africa sometimes reaches 137° Fahr. during the day, and cools sufficiently at night to split off blocks of 200 lbs. weight.

Even in low latitudes and moist climates the effects of temperature changes are often seen. Thin beds of limestone at the bottom of quarries are sometimes so expanded by the heat of the sun as to arch up and break.[1] In desert and arid regions,[2] whatever the altitude, the effects of temperature changes are often striking.

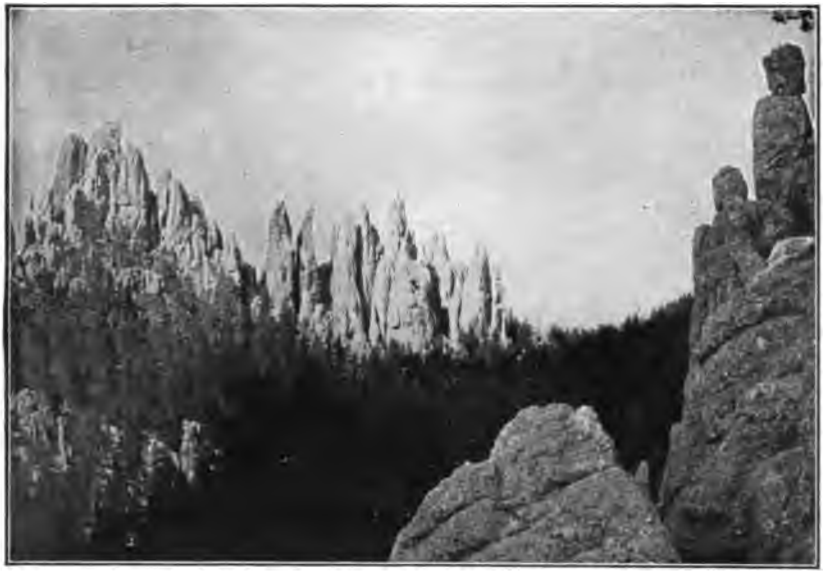

FIG. 32.—Serrate peaks of granitic rock in Black Hills. (Darton, U. S. Geol. Surv.)

The disruption of rock by changes of temperature is one phase of weathering. It tends to the formation of a mantle of rock waste, which, were it not removed, would soon completely cover the solid rock beneath and protect it from further disruption by heating and cooling; but the loose material thus produced becomes an easy prey to running water, so that the work of the atmosphere prepares the way for that of other eroding agencies.

II. Evaporation and Precipitation.

Perhaps the most important work of the atmosphere as a dynamic agent lies in its function as the medium for the circulation and distribution of water. Atmospheric temperature is the primary factor governing evaporation, an important factor in the circulation of the vapor after it is formed, and controls its condensation and precipitation.

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[1] Buckley. Surv. of Wis., Bull. IV, pp. 19, 20.

[2] For an excellent discussion of erosion in dry regions see Walther's Die Denudation in die Wüste.

The average amount of annual precipitation on the land is variously estimated at from forty to sixty inches, the lesser figure being probably more nearly correct. Since much of this water falls at high altitudes, the work which it accomplishes in getting back to the sea is great. The water which falls on the land, if withdrawn wholly from the ocean, would exhaust that body of water in 10,000 to 15,000 years if none of it returned. The work of evaporation is of course not done by the atmosphere, though the atmosphere determines the effect of the solar energy which does the work.[1]

The precipitation is distributed with great inequality, and this inequality affects both the rain and the snow. Some regions have heavy precipitation and some light; some regions have much rain and little snow; others have much snow and little rain; others have rain and no snow, and still others have snow and little or no rain. The amount and distribution of rain and snow determine the size and distribution of streams and glaciers, and streams and glaciers are the most important agencies modifying the surface of the land.

It is impossible to separate sharply the geologic work of the water of the atmosphere from that of other waters; but so long as moisture is in the atmosphere (including the time of its precipitation) its effects are best considered in connection with the atmosphere.

The mechanical work of the rain.—In falling the rain washes the atmosphere, taking from it much of the dust, spores, etc., which the winds have lifted from the surface of the dry land. Not only this, but in passing through the atmosphere the water dissolves some of its gases, and perhaps particles of soluble solid matter. When therefore the falling water reaches the surface of the land it is no longer pure, and some of the gases it has taken up in its descent enable it to dissolve various mineral matters on which pure water has little effect.

As it falls on the surface of the land the rain produces various effects of a mechanical nature. In the first place, it leaves on the surface the solid matter taken from the air. The amount of material, thus added to any given region in any particular shower is trivial, but in the course of long periods of time the total amount of material washed out of the air must be very great.

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[1] On the assumption that condensation takes place at an average elevation of 3000 feet, it has been estimated that the force necessary to evaporate and diffuse the moisture which falls as rain and snow would be equivalent to 300,000,000,000 horse-power constantly in operation. (Strachey, Lectures on Geography, p. 145.)

Every rain-drop strikes a blow. If the drops fall on vegetation, they have little effect, but if they fall on sand or unprotected earthy matter they cause movements of the particles on one another, and this movement involves friction and wear. While the results thus effected are inconsiderable in any brief period of time, they are not so insignificant when the long periods of the earth's history are considered.

Clayey soils contract and often crack on drying. Falling on such a soil when it is dry the rain causes it to expand, and the cracks are healed by lateral swelling. The same soils are baked under the influence of the sun, and when in this condition are softened and made more mobile by the falling of rain. Under the influence of the expansion and contraction occasioned by wetting and drying, the soils and earths on slopes creep slowly downward. When rain falls on dry sand or dust the cohesion is at once increased, and shifting by the wind is temporarily stopped.

After the water has fallen on the land its further work cannot be looked upon as a part of the work of the atmosphere; but any conception of the geological work of the atmosphere which did not recognize the fact that the waters of the land have come through the atmosphere would be inadequate. The work of the water after it has been precipitated from the atmosphere must be considered in another chapter.

III. Effects of Electricity.

Another dynamic effect conditioned by the atmosphere is that produced by lightning. In the aggregate this result is inconsequential; yet instances are known where large bodies of rock have been fractured by a stroke of lightning, and masses many tons in weight have sometimes been moved appreciable distances. Incipient fusion in very limited spots is also known to have been induced by lightning. Where it strikes sand it often fuses the sand for a short distance, and, on cooling, the partially fused material is consolidated, forming a little tube or irregular rod (a *fulgurite*) of partially glassy matter. Fulgurites are usually only a few inches in length, and more commonly

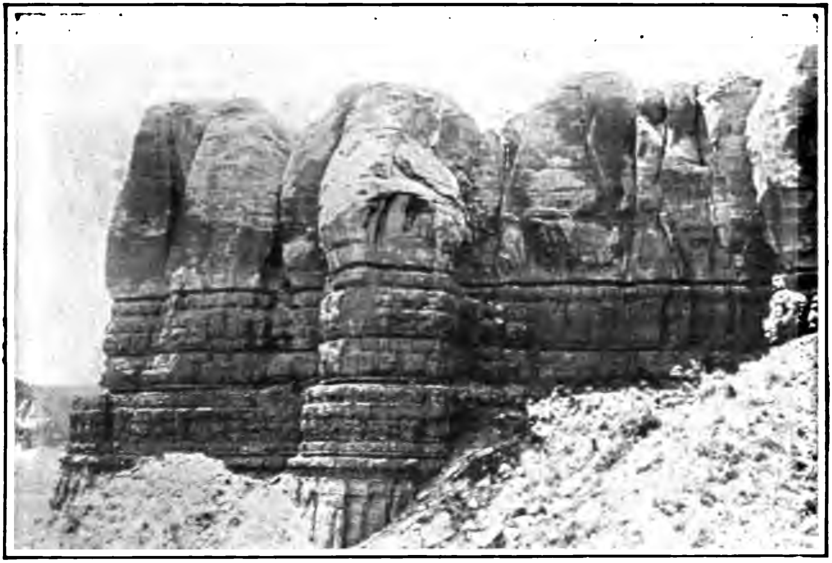

Fig. 33.—Stratified jointed rock in process of weathering. (Cross, U. S. Geol. Surv.)

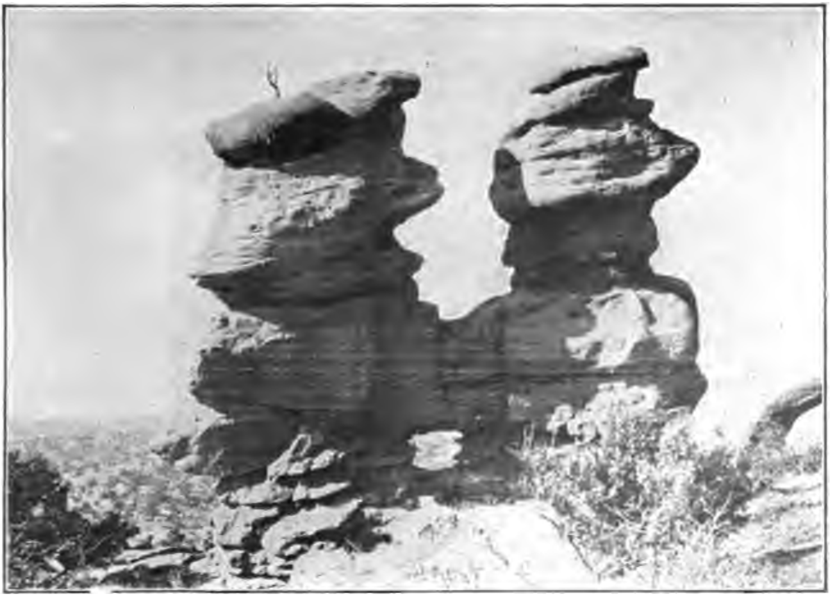

Fig. 34.—Represents a later stage of the processes illustrated by Fig. 33. (Darton, U. S. Geol. Surv.)

than otherwise a fraction of an inch in diameter. Strictly speaking these results are the effect of the electricity of the atmosphere rather than of the atmosphere itself, but they are best mentioned in this connection.

Allusion has already been made to the chemical changes in the atmosphere occasioned by electric discharges.

SUMMARY.

Weathering.—The result of all atmospheric processes, whether physical or chemical, by which surface rock is disrupted, decomposed or in any way loosened, is *weathering*. This convenient term also includes similar results effected by ground water, plants, etc. The tendency of weathering is to produce a mantle of residuary earth over

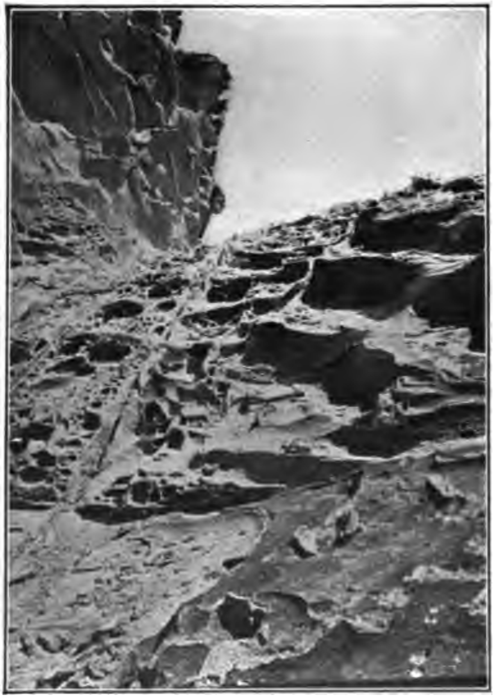

FIG 35.—Details of a weathered rock surface, due partly to wind work and partly to solution. The particular phase of weathering illustrated by this figure is known as "honeycomb" weathering. (Gilbert, U. S. Geol. Surv.)

solid rock. Weathering by mechanical means tends to produce material which, though in a finer state of division, is still like the original rock in chemical composition. Weathering by chemical means tends to produce a mantle made up chiefly of the less soluble parts of the rock from which it was derived. All processes of weathering prepare material for transportation by wind and water.

Many considerations determine the thickness which the mantle of weathered rock (mantle rock) attains. Some of these considerations have to do with the atmosphere, and some with drainage. Since the latter are, on the whole, more important, this matter will be discussed in connection with the work of water (Chapters III and IV).

CHAPTER III.

THE WORK OF RUNNING WATER.

FAMILIAR phenomena, both of land and sea, reveal the constant activity and importance of water as a geologic agent. Even when there is no precipitation the moisture in the air influences its activity in certain ways. Just as iron "rusts" more readily in moist air than in dry, so changes in other mineral substances are influenced by atmospheric humidity. Where precipitation takes place the results are more obvious. The passing shower works changes in the surface of the land, striking in proportion to the rate and amount of precipitation. The rains feed the streams, and every stream is modifying its bed, and with increasing rapidity as its current is swollen. Even the moisture which is precipitated as snow works its appropriate results. Before it melts it protects the surface against other agents of change; but if it accumulates in sufficient quantity in appropriate situations, it may give rise to avalanches and glaciers, which, like running water, degrade the surface over which they pass.

A part of the water which falls as rain, and a part of that which results from the melting of snow and ice, sinks into the soil and into the rock below, becoming *ground water*. It is this ground water which especially justifies the name *hydrosphere*, often applied to the waters of the earth, for it literally forms a spherical layer in the outer portion of the solid part of the earth. During the stay of the water beneath the surface it effects changes in the rocks through which it passes, dissolving mineral matter here and depositing it there, substituting one substance for another in this place, and effecting new chemical combinations in that. Slow as these processes are, they have worked wondrous changes in the course of the earth's history.

When the waters are gathered together in ponds, lakes, and oceans, they are still active, and the results of their activity are seen along the shores, where winds and waves produce their chiefest effects. Even the ocean currents, far from land, and the processes of the deep sea, are not without their effect on the course of geological history.

The work of the *surface waters*, *ground* (*underground*) waters, *standing waters*, and *ice* will be considered in order.

RAIN AND RIVER EROSION.

Rain and river erosion began when the first rains fell on land surfaces. Neither the location nor the nature of the first land surface is known. There is little reason to believe that the ocean was ever universal, but there is reason to believe that most land areas have at some time or other been covered by the sea. The prevalent conception that land areas which were once submerged came into existence by being elevated above sea-level, should be supplemented by the alternative conception that submerged areas may have become land by the depression of the ocean basins, thus drawing off the water from the areas where it was shallow. Thus in Fig. 36 the sinking of the sea-bottom from *a* to *b* would lower the surface of the water from *cc'*, to *dd'*, and draw off the water from the surfaces *cd* and *c'd'*.

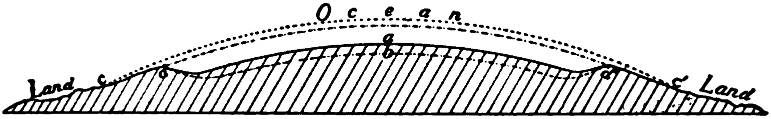

Fig. 36.—Diagram to illustrate the origin of lands by the lowering of the sea-level due to depression of the sea bottom. If the bottom is depressed from *a* to *b* the surface will be drawn down from *cc'* to *dd'*, and the surfaces *cd* and *c'd'* will become land.

Without attempting to picture the character of the original land our study of subaërial erosion may begin with an area which has just been changed from sea bottom to land. What is the nature of such a land surface? Of what material is it composed, and what is the character of its topography? Concerning its constitution something may be inferred from the nature of the deposits now found at the bottom of the sea. Near the shore and in shallow water they often consist of gravel and sand, though other materials are not wanting. Far from shore and in deep water they consist for the most part of fine sediments, some of which were washed or blown from the land, some of which came from the shells and other secretions of marine animals, some from volcanoes, and some from various other sources. The topography of the newly emerged land may have had some likeness to the topography of the sea bottom. The numerous soundings which have been made over large areas of the sea have shown that its bottom is, as a rule, free from the numerous small irregularities which affect the surface of the land. They seem to show that a large part of the ocean bottom is so nearly flat that, if the water were removed, the eye would hardly detect irregularities in the surface. This statement does not lose sight of the fact that the ocean bottom is, in certain places. markedly irregular. Volcanic peaks and striking irregularities of other sorts abound in some places. Nevertheless if the bottom of the sea could be seen as the land is, its most striking feature, taken as a whole, would be its apparent flatness.

With the topography of the sea bottom the topography of the land is, in its details, in sharp contrast. In order to get at the history of the latter, we may study the sequence of events which would follow the emergence of a portion of the former.

Subaërial Erosion without Valleys.

For the sake of emphasizing the fundamental principles involved in the work of running water, a hypothetical case will first be studied in some detail, even at the risk of elaborating processes already understood. The principles themselves will find application later in relations which are much less simple.

Let it be assumed that the area of newly emerged land is a circular dome-shaped island. The simplest possible condition is represented by assuming its slope to be the same in all directions from the center, and its materials to be absolutely homogeneous. Such an island would be subject to all the forces ordinarily operating on land surfaces. The chief agency tending to modify land surfaces is atmospheric precipitation. It will be assumed that the rain falls on the surface of the island with absolute equality at all points, and that all other forces which affect it operate equally everywhere.

The rain falling on a land area disappears in various ways; part of it evaporates, part of it sinks, and part of it runs off over the surface. If the island be composed of fine and unconsolidated materials, such as clay, the water which runs off over the surface will carry sediment down to the sea. If the island be composed of solid rock instead, exposure to the air will cause it to decay, and the products of decay, such as sand and mud, will suffer a like fate.

For the sake of a clear understanding of the processes involved, two cases may be postulated; one in which the waters of the sea remove the sediment washed down from the hypothetical island as fast as it reaches the shore, and one in which they allow it to accumulate without let or hindrance. In both cases the wear of the waves will be neglected.

1. In the first case the water flowing off over the surface (the *run-off*) will descend equally in all directions. It will constitute a continuous sheet of surface-water, and both its volume and its velocity will be the same at all points equally distant from the summit. Erosion accomplished by sheets of running water, as distinct from streams, is *sheet* (or *sheet-flood*) *erosion*.[1] Since the material of the surface is homogeneous, the wear effected by the water will be equal at all points where its velocity and volume are equal. For obvious reasons the depth of the run-off will increase from summit to base. The gradient (slope) also increases in the same direction, and the increase of volume and of gradient conspire to augment the velocity of the water, and therefore of the wear effected by it. If the thin sheet of water starting from the top of the island with relatively low velocity be able to wash off even a little fine material from the surface, the thicker sheet farther down the slope, moving with greater velocity, will be able to carry away more of the same sort of material, and the increase will be progressive from summit to base. It follows, therefore, that the surface will be worn equally at points equally distant from the summit, but unequally at points unequally distant from it. The first shower which falls on the island may be conceived to wash off from its surface a very thin sheet of material, but a sheet which increases in thickness from top to bottom. The run-off will not be stopped immediately on reaching the sea, but will displace the sea-water to some slight depth, and wear the surface some trivial distance below the normal level of the sea. The result of successive showers working in the same way through a long period of time will be to diminish the area of the island and to steepen its slopes. The results of a considerable period of erosion under these conditions are shown

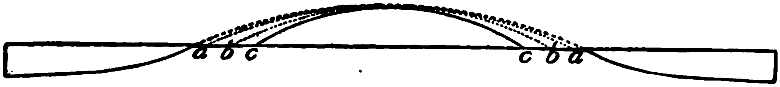

Fig. 37.—Diagram to illustrate the effect of rain erosion on an island where there is no deposition or wave erosion about its borders. The uppermost curve represents the original surface, while *aa*, *bb*, and *cc* represent successive surfaces developed by sheet erosion, on the supposition that no material is deposited along the shores.

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[1] McGee. Bull. Geol. Soc. Am., Vol. VIII, pp. 87–112.

diagrammatically in Fig. 37, which illustrates both the diminution in area which the island has suffered, and the increase in the angle of its slopes. Immediately about it, at the stage represented by *aa*, Fig. 37, there is a narrow marginal platform, or submerged terrace, in place of the land area which has been worn away at or just below the level of the sea. Long successions of rains working in the same way will give the island steeper slopes, a smaller area, and a wider marginal terrace. Successive stages are shown by the lines *bb* and *cc*, Fig. 37.

If rain falls on such an island until it completes the work which it is possible for running water to do, the island will be reduced essentially to the level of the sea, and in its place there will be a plain, the area of which will be equal to that of the original island. Its central point will be at the level of the sea, and its borders a trifling distance below it (Fig. 38). The island is gone, and in its place there is a plain as low as running water can wear it. Other agencies might come in to defeat the result just outlined, but if the island did not rise or sink after its formation, rain falling upon it would, *under the conditions specified*, finally bring about the result which has been sketched. The plain (Fig. 38) which succeeds the island is a *base-level of erosion*, though this term is also used in other ways. Under these conditions the slope of the land would remain convex at all stages, but the convex erosion profile of the land would meet a nearly straight line just below sea-level. The relative lengths of these two elements of the profile, the curve above and the straight line below, vary as erosion progresses, the convex portion becoming shorter and the other longer. The two parts of the profile taken together are concave upward at the lower end all the time, and for a greater distance from its lower end in all the advanced stages of erosion (Fig. 37).

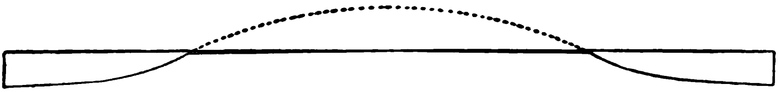
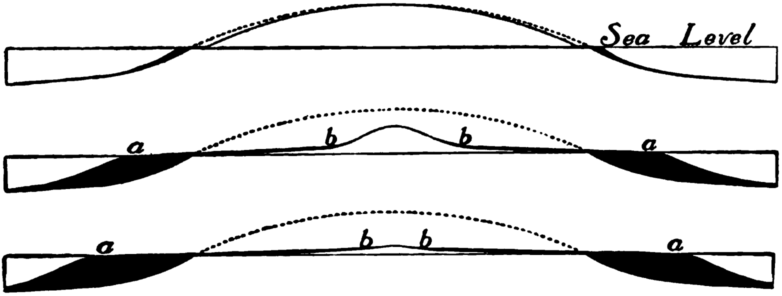


Fig. 38.—Diagram to illustrate the final effect of rain erosion under the conditions specified in the text. The diagram expresses the final result of the processes suggested by Fig. 37.

In the destruction of the land under these conditions *neither valleys nor hills would be developed*, nor would the topography of the land be fashioned to correspond with the surfaces with which we are familiar.

It is to be distinctly borne in mind that the foregoing is a hypothetical case; it is not probable that such an island ever existed, or ever will; but that does not diminish the value of the illustration, since the principles involved are operating on every land mass, though in less simple relations.

2. The second case differs from the first in that the sediment washed down from the land is deposited about its borders. This results in the building up of a marginal platform, as shown in Figs. 39–41. As erosion goes on more sediment is washed down and deposited, partly on the narrow marginal shelf which has already been developed, and partly on its outer slope, as shown in the figures. The marginal flat is thus extended beyond the original shores of the island on the one hand, and toward its center on the other. As it develops, its inner portion, and indeed all except its outer edge (*ab*, Figs. 40 and 41), will be gradually built up above the level of the water. This marginal lowland is developed at a level as low as running water, *under the conditions then and there present*, can reduce the land. Such a surface may be said to be *at grade*, since running water neither wears it down nor builds it up. Its angle of slope is a function of (1) the volume of the water running over it, and (2) of the load which the water carries.



Figs. 39–41.—Diagrams to illustrate the effect of rain erosion on an island when all the eroded material is deposited about the shore. The black portions represent deposition The dotted lines represent the original surface The several diagrams represent successive stages in the process.

Since the marginal plain of the above illustration extends beyond the original shore of the island, the area of land is increased, though both its average elevation and its mass (above water) are reduced. In case destructive processes did not operate on the marginal graded plain the spreading and lowering suggested by Figs. 39 and 40 would go on until the central mass of the island was brought down to a gradient in harmony with that of the gently sloping border, as shown in Fig. 41. When this had been accomplished there would be a relatively large land area with low slopes (Fig. 41) in place of the smaller area with steeper ones (compare Figs. 39 and 40). The basal part of the larger island from the center to the original margin would be made up of the original material in its original position (unshaded part of Fig. 41). Its surface would be covered, least deeply near its center and most deeply near the original margin, with débris gradually shifted from higher levels, as shown in Fig. 41.

Were such an island as that shown in Fig. 41 once formed, the rain falling on it, and flowing off over its surface, would carry off its surface soil and spread it about the shores. Though the surface of the marginal flat of Fig. 40 was as low as running water could bring it at the time it was developed, the conditions of erosion have changed by the time the land reaches the conditions shown in Fig. 41, and the same amount of rainfall may now be effective in erosion. In the first case (Fig. 40) the water descending from the higher part of the land brought down sediment and started across the flat with a load. Its energy was consumed in transporting what it had, not in getting new material. In the second case (Fig. 41) the water flowing over the gently sloping surface has no initial load, and its energy is therefore available for erosion. Under continued rainfall, the area of the land shown in Fig. 41 would be increased as before by successive marginal deposits (see Fig. 42), and at the same time its average height would be reduced. The lowering and enlarging of the island would continue until the whole surface was brought so nearly to the level of the sea that water would cease to run over it with sufficient velocity to carry away even the fine material of its surface. Such a surface, brought down as low as running water can degrade it, is also (see p. 57) a *base-level.* It will be seen from the foregoing illustrations that a *graded surface* may pass into a *base-level,* with no sharper line of demarkation

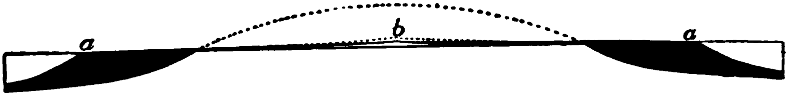


FIG. 42.—Diagram to illustrate the result of the continuation of the processes shown in Figs. 39–41.

than that which separates a mature man from an old one. In this case, as in the preceding, the island has been base-leveled, but still without the formation of valleys or hills.

Both the preceding hypothetical cases make it clear that, from the point of view of erosion, every drop of water which runs off over the surface of the land has for its mission the getting of the land into the sea. Under ordinary conditions surface drainage must fail to bring a land area altogether to sea-level, the absolute base-level of subaërial forces; but it is not simply the water which runs off over the surface which degrades the land. That which sinks beneath the surface contributes to the same end by slowly dissolving mineral matter below the surface, and finally carrying it to the sea. In this way the reduction of land areas to sea-level may be completed.

The rain-water which evaporates from the surface without sinking beneath it does not effect much wear; but the water thus evaporated is subject to re-precipitation, so that, in the long run, it may assist in the work which has been sketched. Thus it is not simply the waters which run off over the surface of the land, but all which fall upon it, which unite to compass its destruction.

The Development of Valleys.

By the growth of gullies.—Had the slopes of the hypothetical island not been absolutely uniform the processes of erosion would have been different. Let the departure from uniformity be supposed to consist of a single slight meridional depression near the base of the island (Fig. 43). As the rain falls it will no longer run off equally in all directions. A greater volume will flow through the depression than over other parts of the surface having the same altitude, and the greater volume of water along this line will give greater velocity, greater velocity will occasion greater erosion, and greater erosion will deepen the depression. The immediate result is a *gully* or *wash* (Fig. 44). So soon as the gully is started it tends still further to concentrate drainage in itself, and is thereby enlarged. The water which enters it

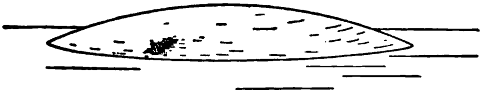


FIG 43.—Diagram showing a slight meridional depression in the surface of an otherwise even-sloped island.

from the sides widens it; that which enters at its head lengthens it by causing its upper end to recede; and all which flows through it, so long as its bottom is above base-level, deepens it. The enlarged gully will gather more water to itself, and, as before, increased volume means increased velocity, and increased velocity increased erosion. As the gully grows, therefore, its increased size becomes the occasion of still further enlargement.

Continued growth transforms the gully into a ravine, though between a gully and a ravine there is no distinct line of demarkation. But growth does not stop with ravine-hood. Water from every shower gathers in the ravine, and, flowing through it, increases its length, width, and depth, until it reaches such proportions that the term *ravine* is laid aside, as childhood names are, and the depression becomes a valley.

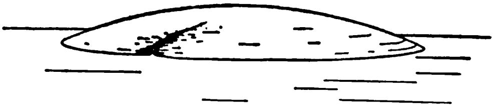


Fig. 44.—Diagram illustrating the development of a gully, starting from the conditions shown in Fig. 43.

It was assumed in the preceding paragraphs that the single depression in the slope was meridional and low on the slope, but almost any sort of depression in almost any position would bring about a similar result, since it would lead to concentration of the run-off. Had the original surface been interrupted by ridges instead of depressions, the effect on valley development would have been much the same, for a ridge, like a depression, would, in almost any position, occasion the concentration of the run-off, and so the development of valleys. Under the conditions represented in Fig. 44 the lengthening of the drainage depression is effected chiefly at its upper end, the head of the valley working its way farther and farther back into the land. This method of elongation is known as *head erosion.* But the lengthening of the valley is not always wholly by head erosion. The gully normally begins where concentration of run-off begins, and if this were not at sea-level, the gully might be lengthening at both ends at the same time. This would have been the case, for example, had the original depression of Fig. 43 been half-way up the slope of the island.

If while the slopes of the island were absolutely uniform its surface material failed of homogeneity, the result would be much the same as if the slopes were unequal. If the material lying along a certain meridian of the island be slightly softer than that over the rest of the surface, the run-off, which would at the outset be equal on all sides, would effect more erosion along the line of the less resistant material than elsewhere. The result would be a depression along this line, and, once started, the depression would be a cause of its own growth. If the soft material were disposed in any way other than that indicated, the final result would be much the same, for it would quickly give origin to a depression which would lead to the concentration of the surface-waters, and this is the condition for the development of a gully, a ravine, and finally a valley.

In the presence of sufficient rainfall, either heterogeneity of slope or of material will therefore occasion the development of valleys. If the lack of uniformity appears at but a single point there will be but a single valley. If it appears at many points the number of valleys will be large. Since it is incredible that a land mass of perfectly homogeneous material and of absolutely uniform slopes ever existed, it is believed that every land mass, affected for any considerable length of time by rain, has had valleys developed in it. The degree of heterogeneity of material and slope is usually so great as to lead to the development of many valleys, even on areas which are not large; but for the sake of emphazing the simpler elements of the complex processes of stream work, the hypothetical case of an island with but a single valley, and that without tributaries, may first be studied. Under these conditions two cases may be considered, the one where there is no deposition about the island, and the other where deposition takes place.

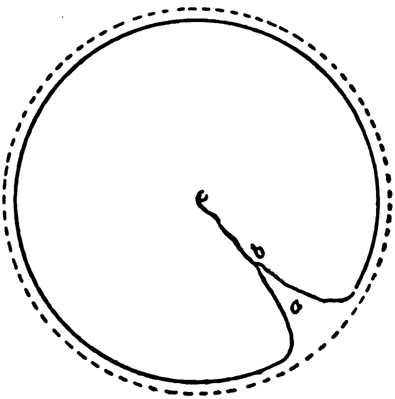


FIG. 45.—Diagram to illustrate the effect of sheet and stream erosion on the outline of an island when no deposition takes place about its borders. The dotted line represents the original outline of the island, the full line its border at a later time. The stream develops a reëntrant (bay) in the outline.

1. If all the material eroded from the surface of such an island, both in and out of the valley, were carried well beyond the borders of the land before being deposited, the edge of the island would re-

cede from its original position toward the center, as illustrated by Figs. 37 and 45; but the recession would be most rapid where the valley joins the sea (Fig. 45). At this point therefore a reentrant would be developed (*a*, Fig. 45), and the island would lose its circular outline. Continued erosion would cause the shore-line to retreat on all sides, but fastest at the lower end of the valley, and the final result would be a base-level differing from that developed under the conditions specified on p. 60, in that the last part to be brought low would not be the center of the original island.

Under the foregoing conditions the profile of that part of the valley which is above sea-level (*cb*) would be convex, following the analogy of sheet erosion on a hypothetical island of uniform slopes and homogeneous material with no marginal deposition. Its side slopes, likewise developed under the influence of running water augmented in volume from top to bottom, would also be convex.

2. If the sediment washed down from the land is deposited about its borders, both the outline of the island and the profile of the valley will be altered. Deposition at the debouchure of the valley follows the same principles as deposition elsewhere; but if all the sediment brought to the sea be deposited at the shore, the seaward extension of the land by deposition would be more rapid opposite the valley than elsewhere, and the island would lose its circular outline, and develop some such form as is shown in Fig. 46. In this case the profile of the upper end of the valley, and the upper parts of its side slopes, as well as the upper parts of the extra-valley slopes of the island, are convex (compare Figs. 39 and 40); but the convexity above is exchanged for concavity below, the change beginning at the point where downward erosion of the descending waters is checked. As a valley lengthens, the larger part of its

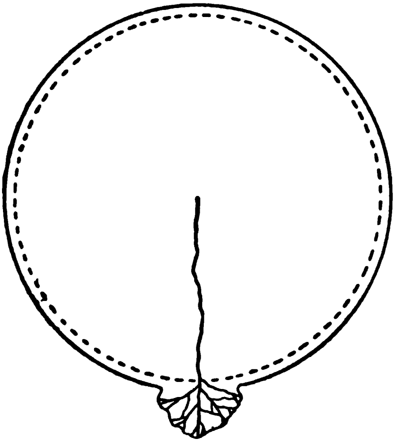


FIG. 46.—Diagram showing the outline of an island as modified by sheet and stream erosion where eroded material is deposited at the shore. The dotted line represents the original outline; the full line, a later one. The excess of deposition at the end of the valley causes a projection of land into the sea.

profile becomes concave (compare the profiles of Figs. 39 to 41), but the extreme upper end still remains convex. Since the side slopes of a valley are much shorter than its lengthwise slope, a larger proportion remains convex. Under the conditions here discussed the change from convexity above to concavity below would begin at about the point where deposition begins.[1]

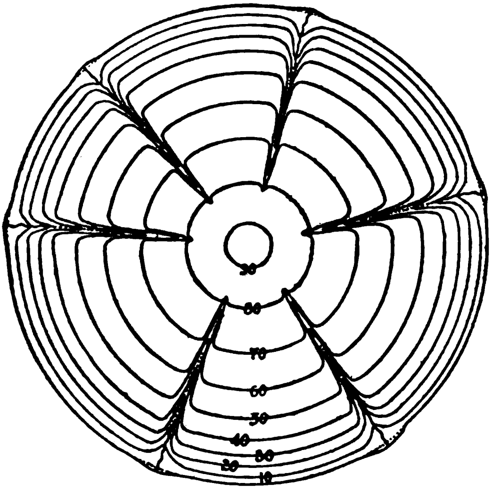


FIG. 47.—Diagram representing several meridional valleys developing in a circular island. The valleys are all young and narrow. All are making deposits at their debouchures.

The deposition at the debouchure of the valley, and later above the debouchure, will follow the same course as about the island under the conditions already discussed (pp. 61, 62).

Limits of growth.—In all cases there are limits in depth, length, and width, beyond which a valley may not grow. In depth it may reach base-level. At the coast, base-level is sea-level,[2] but inland it rises by a gentle gradient. In length, the valley will grow as long as its head continues to work inland. In the case represented by Figs. 45 and 46 the head of the valley would not cease to advance when the center of the island was reached, though beyond that point head erosion would not be more rapid than lateral erosion on either side. If but a single valley affected a land area the limit toward which it would tend, and beyond which it could never pass, would be the length of the land area in the direction of the valley's axis. In width, a valley is increased

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[1] For a discussion of convex and concave erosion slopes see Bain, Geol. Surv. of Ia., Vol. VI, p. 449.

[2] Great rivers, like the Mississippi, cut their *channels* somewhat below sea-level, but probably not by an amount exceeding the depth of the stream itself (see p. 79).

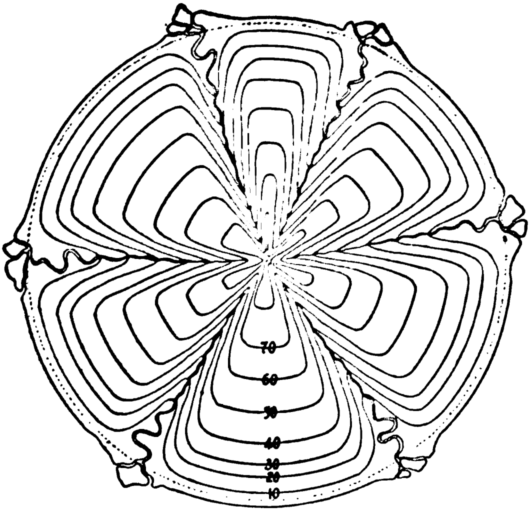

FIG. 48.—Same as Fig. 47, with the valleys more developed. The dotted line represents the original outline of the island. Its area is being extended by deposition everywhere, but most at the debouchures of the streams.

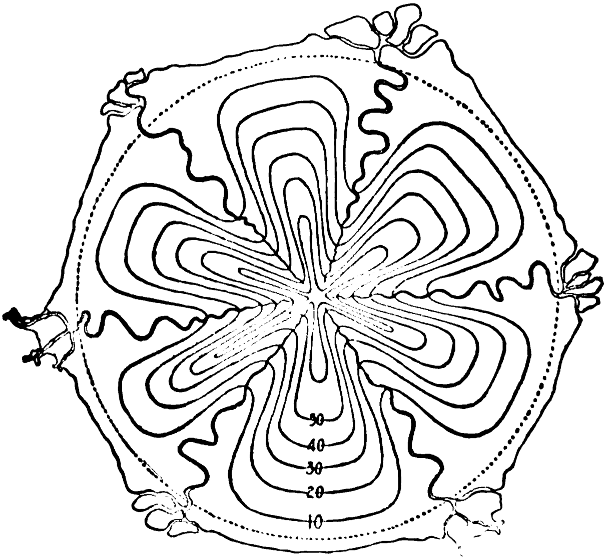

FIG. 49.—A later stage of the island shown in Fig. 48.

by the side cutting of the stream, by the wash of the rain which falls on its slopes, and by the action of gravity which tends to carry down to the bottom of the slope the material which is loosened above by any process whatsoever. If there be but one valley in a land area its limiting width is scarcely less than the width of the land itself.

Had there been several initial meridional depressions instead of one in the island, or had there been several meridional belts where the material of the surface was less resistant than elsewhere, several valleys would have been developed, converging toward the center (Fig. 47). If the conditions were such as to allow of the equal development of valleys on all sides of the island, each would be lengthened by head erosion until it reached the center of the island, where the permanent divide between their heads would be established. Each would be widened by all the processes which widen valleys, and their widening would narrow the intervening areas (Figs. 48 and 49). Under conditions of equal erosion the limits of width for each valley would be the centers of the ridges on either side, and here the divides between them would be permanently established. Though erosion would continue even after the crest of the ridge had been narrowed to a line, the permanence of the divide would follow from the fact that erosion would be equal on both sides of this line, and its effect would be to lower the divide, but not to shift it horizontally (Fig. 50). The limits in length and width are therefore not the same where there are several valleys as where there is but one. The limit in depth, however, remains the same, and the final result of erosion, proceeding along these lines, would be the base-leveling of the land, leaving a plain but slightly above sea-level. The plain would not be absolutely flat, though its relief would be very slight, and the higher parts would be along the lines of the divides between the streams (Fig. 51. Compare also Fig. 42). Many valleys would occasion more rapid degradation than few, and the period of base-leveling would be correspondingly shortened.

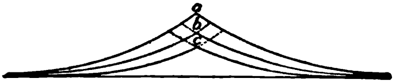

FIG. 50.—Diagram to illustrate the lowering of a divide without shifting it. The crest of the divide is at *a*, *b*, and *c* successively. If erosion were unequal on the two sides, the divide would be shifted

Had the initial depressions which gave origin to the valleys had positions other than meridional, the valleys would have had other and less regularly radial courses, but the final result of their development would have been the same.

It is not to be inferred that the method of valley development which has been sketched is the only one. The processes of valley development are complex, and the history of some valleys has run a different course; yet the processes outlined above are in operation in all cases, and they were probably the most important ones in the development of the first drainage system on any land surface. As will be seen in the sequel the history of valleys is subject to serious accidents, and they are often of such a nature as to mask the simplicity of the more normal processes.

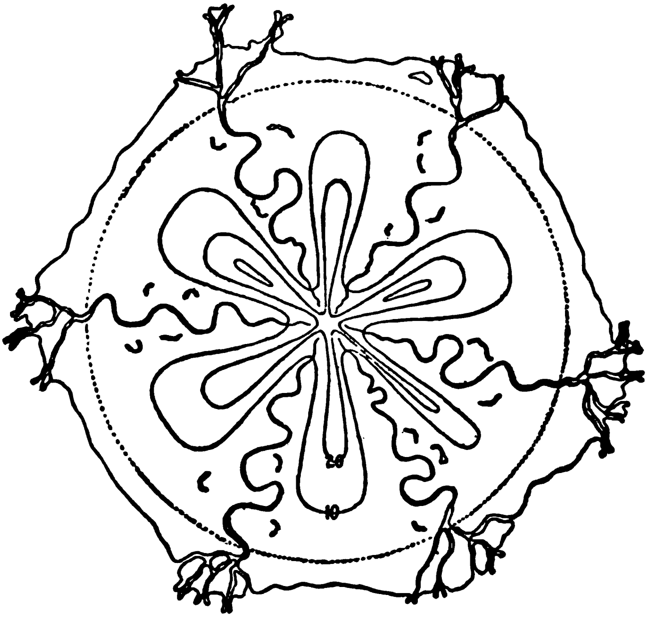

FIG. 51.—Diagram illustrating the further development of Fig. 49. The land here has been reduced greatly, though not yet to base-level.

The permanent stream.—From the foregoing discussion, it is seen that a valley may be developed by the run-off of successive showers. If supplied only from this source surface streams would cease to flow soon after the rain ceased to fall, and a valley might attain considerable size without possessing a permanent stream. How does the valley developed by the run-off of successive showers come to have a permanent stream? The answer to this question involves a brief consideration of that part of the rainfall which sinks beneath the surface.

If wells be sunk in a flat region of uniform structure and composition the water in them is generally found to stand at a nearly common level. The meaning of this fact is not far to seek. If a hole 60 feet deep fills with water up to a point 20 feet from the surface, it is because the material in which the well is sunk is full of water up to that level. When the well is dug the water leaks into it, filling the hole up to the level to which the rock (or subsoil) is itself full. This level, below which the rock and subsoil (down to unknown depths) are full of water, is known as the *ground-water level, ground-water surface,* or *water-table.*

The ground-water level fluctuates. In a wet season it rises, because more water has fallen and sunk beneath the soil; but several processes at once conspire to bring it down again. Where there is growing vegetation its roots draw up water from beneath, and evaporation also goes on independently of vegetation. The water is drawn out through wells and runs out through openings. It may also flow underground from one region to another where the ground-water surface is lower. All these processes depress the ground-water surface.

A well sunk to such a level as to be supplied with abundant water in a wet season may go dry during a period of drought because the ground-water level is depressed below its bottom. Thus either well shown in Fig. 52 will have water during a wet season when the water-level is at *a*; but well No. 1 will go dry when the water surface is depressed to *b*.

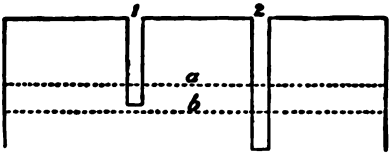

Fig. 52.—Diagram illustrating the fluctuation of a ground-water surface. *a* = wet-weather ground-water level; *b* = ground-water level during drought. Well No. 1 will contain water during the wet season, but will go dry in times of drought. Well No. 2 will be permanent.

The principles applicable to wells are applicable to valleys. When a valley has been deepened until its bottom reaches below the ground-water level, water seeps or flows into it from the sides. The valley is then no longer dependent on the run-off of showers for a stream. It will be readily seen that at some stage in its development, the bottom of a valley may be below the ground-water level of a wet season without being below that of a dry one. Thus the valley represented in cross-section by the line 2–2, in Fig. 53, will have a stream when the ground-water level is at *aa*, but none when this level is depressed to *bb*. If the rainfall of the year were concentrated in a single wet season, the intermittent stream would flow not only during that season, but for so long a time afterward as the ground-water level remained well above the valley bottom. In regions subject to frequent and short periods of heavy precipitation, alternating with droughts, the periods of intermittent flow may be many and short. Since the precipitation of many regions varies greatly from year to year, it follows that a stream may flow continuously one year and be intermittent the next. Many valleys in various parts of the earth are now in the stage of development where their streams are intermittent.

As a valley containing an intermittent stream becomes deeper, the periods when it is dry become shorter, and when it has been sunk below the lowest ground-water level, it will have a permanent stream (3, Fig. 53).

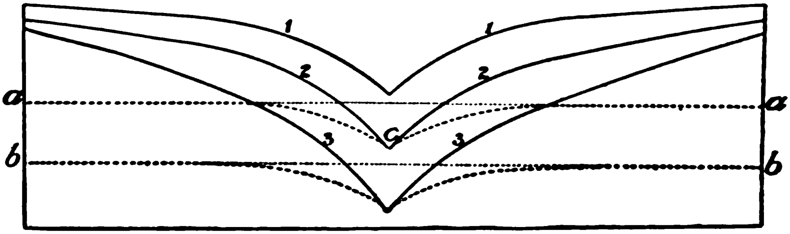

Fig. 53.—Diagram to illustrate the intermittency of streams due to fluctuations of the ground-water level. The water level *aa* would be depressed next the valley 2–2 by the flow of the water into the valley. The profile of the ground-water surface would therefore be *aca* rather than *aa*.

Since a valley normally develops headward, its lower and older portion is likely to acquire a permanent stream, while its upper and younger part has only an intermittent one (Fig. 47 and Fig. 1, Pl. III, near Anthony, Kan. The intermittent part of the stream is indicated by the dotted blue line). For the same reason the head of a stream is likely to be farther up the valley in wet weather than in dry. So soon as a valley gets a permanent stream, the process of enlargement goes on without the interruption to which it was subject when the supply of water was intermittent.

In general a permanent stream at one point in a valley means a continuous stream from that point to the sea or lake which the valley joins; but to this rule there are many exceptions. They are likely to arise where a stream heads in a region of abundant precipitation, and flows thence through an arid tract where the ground-water level is low, and evaporation great. In such cases, evaporation and absorption may dissipate the water gathered above, and the stream disappears (Fig. 2, Pl. III, near Paradise, Nev.).

PLATE III.

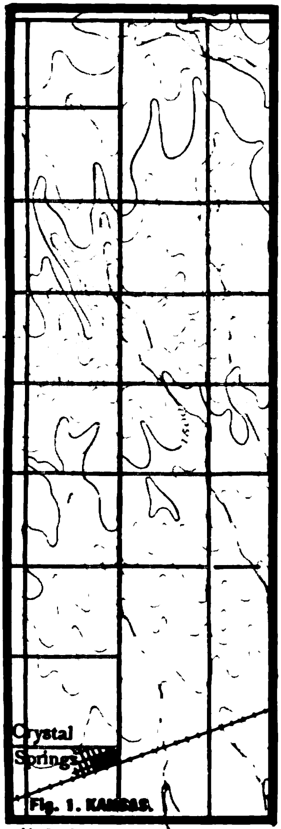

Fig. 1. KANSAS.

U. S. Geol. Surv.

Scale, 2+ miles per inch.

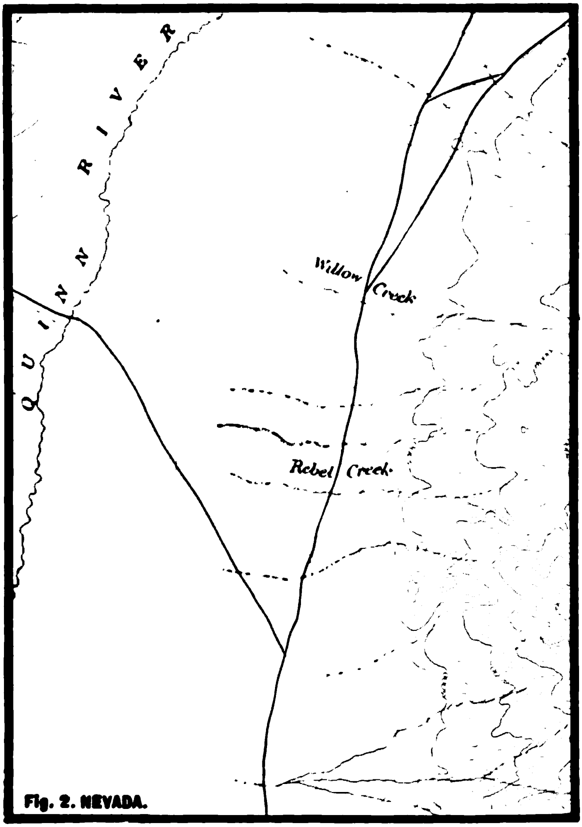

Fig. 2. NEVADA.

U. S. Geol. Surv.

Scale, 4+ miles per inch.

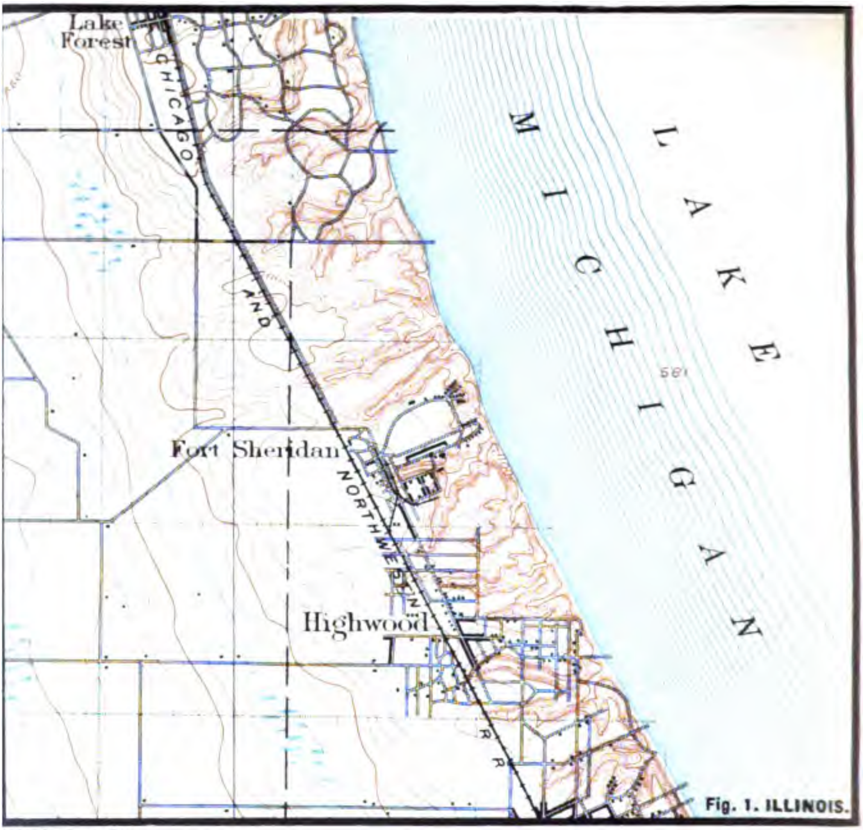

Fig. 1. ILLINOIS.

U. S. Geol. Surv.

Scale, 1+ mile per inch.

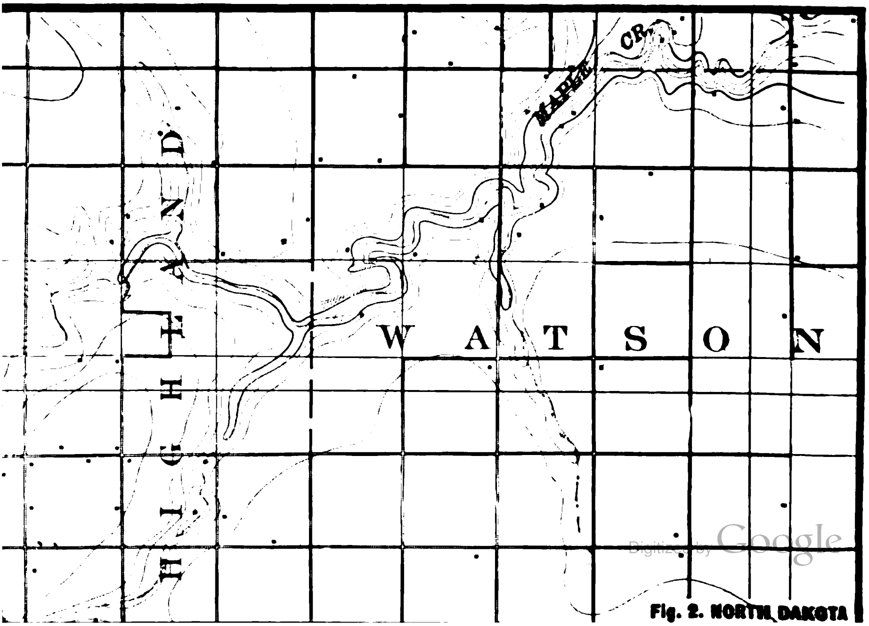

Fig. 2. NORTH DAKOTA

Other modes of valley development.—If as a new area of land emerges from the sea its surface has a depression without an outlet, and such an assumption is by no means improbable, the depression would be filled with sea-water. The inflowing water from the surrounding land might fill the basin to overflowing, and the outflow, finding exit at the lowest point in the rim of the basin, would flow thence toward the sea. Such a stream would develop a valley, the history of which would be somewhat different from that which has been sketched. Instead of developing headward from the sea, the valley would be in process of excavation all the way from the initial basin to the sea at the same time (Fig. 54). The upper end of the valley might ultimately be cut to the level of the bottom of the basin, when the lake would disappear. The head of the

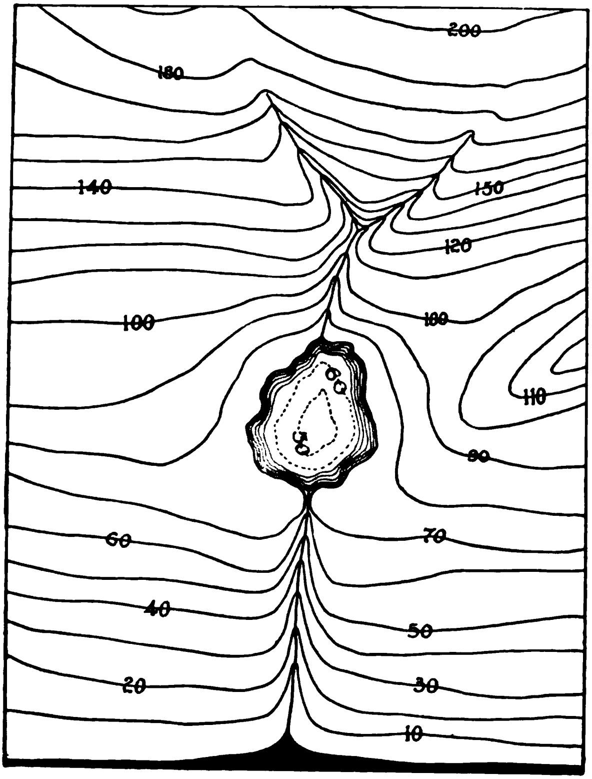

Fig. 54.—Diagram to show how a valley may be developing all the way from a water-filled basin (lake) to the sea at the same time. Small valleys leading to the lake are also developing. The black area = the sea.

valley might then work back across the former site of the lake into the territory beyond. Valleys might have developed above the lake before it was drained, and after this event, such valleys would make connections with the valley below (Fig. 55). A valley developed in this manner is not simply a gully grown big by head erosion, and the valley would not precede the stream.

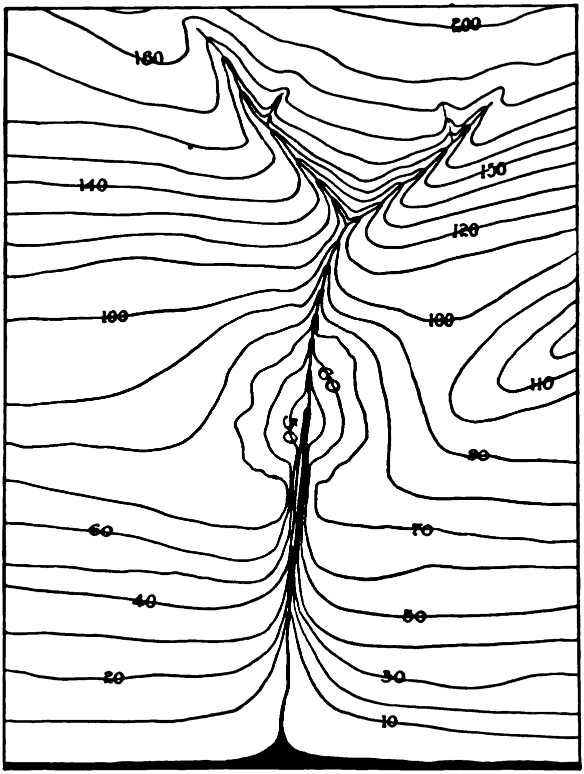

FIG. 55.—The stream leading out from the lake (Fig. 54) has drained the lake, and the valleys above and below the site of the former lake have united.

If a surface of land were notably irregular before valleys were developed in it, there might be many lakes, and the flow from a higher lake might pass to a lower. If the lakes were ultimately drained, the several sections of the valley would be joined to one another without intervening basins. In certain regions, especially those which have been affected by continental ice-sheets, this has been a common method of valley development in *post-glacial* time. In this case also the stream precedes its valley, and not the valley its stream. Many post-glacial valleys, on the other hand, antedated their permanent streams, as in the cases first described.

If the gradient of a slope on which valleys are to develop is notably unequal, though without basins, the development of valleys may follow somewhat different lines. If on emergence the seaward part of a new land area assumes the form of a plain, bordered landward by a steeper slope (Fig. 56), the most notable early growth of the valleys would be on the latter. The run-off would develop gullies on the steep slope, but on reaching the plain below the velocity of the water would be checked, and it would drop much of the detritus washed down from above. This deposition would build up (aggrade) the surface, and much or even all the water might sink into and seep through the débris thus deposited, and disappear altogether from the surface, as at *b*, Fig. 56. This would be most likely to occur where the débris is abundant and coarse, and the precipitation slight. If the water disappears at the base of the mountain (see Fig. 2, Pl. III), the early growth of the valley may be confined to

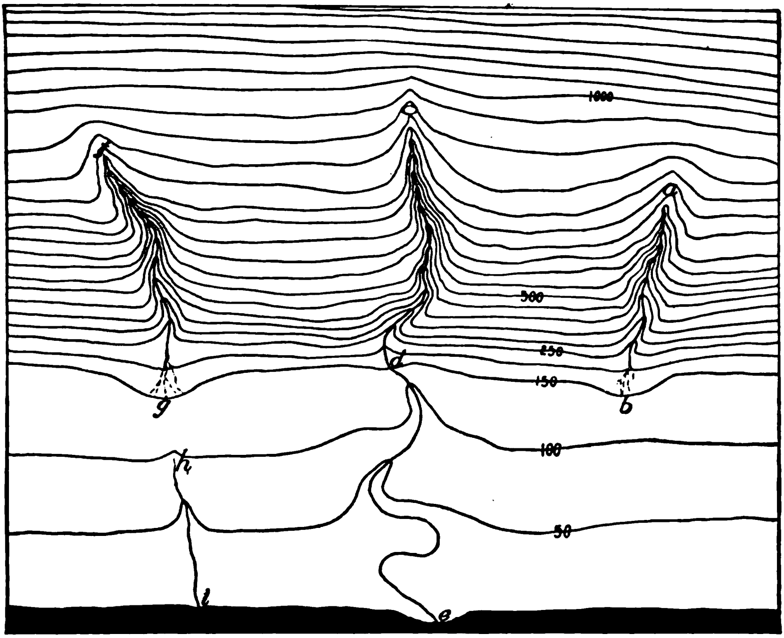

FIG. 56.—Diagram showing the phases of valley development described in the text.

the steep slope remote from the sea (*ab*, Fig. 56); but on the slope where the valley is growing there will be headward lengthening, as in the general case already considered. If the surface drainage does not disappear at the base of the steep slope, the run-off will find its way over the plain along the lowest accessible route to the sea (*de*, Fig. 56). In this case the valley may be growing throughout its length at the same time.

The conditions represented by *ab*, Fig. 56, may be no more than temporary. Sooner or later a valley developing headward across the plain (*hi*, Fig. 56) may provide a channel for the water descending from the higher land beyond. In this case the valley develops in sections, the one on the slope above, the other on the plain below, and their union (compare *fghi*, Fig. 56, with Fig. 57) results from their growth.

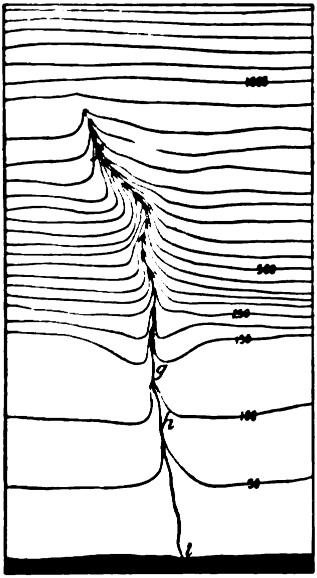

FIG. 57.—Diagram representing the further development of the valleys *fg* and *hi* in Fig. 56. The head of the latter (Fig. 56) has worked back until it has reached the lower end of the former.

The principles here sketched have been in operation wherever land areas were so elevated as to give rise to unequal slopes, and this has perhaps been the rule rather than the exception. The results effected by the operation of these principles would of course be dependent on the varieties of slope, on the abruptness with which a slope of one gradient gave place to another, on the texture of the rock, the amount and distribution of precipitation, etc., etc.

In the preceding paragraphs the lengthening of a valley at its upper end by head erosion has been repeatedly referred to. If all valleys began their development at the sea and lengthened headward, it might seem that their seaward ends should be their oldest parts; but since the development of valleys is begun somewhat promptly after the land appears above the sea, and since the emergence is generally gradual, that part of a valley which is at the seashore at one time may be far inland a little later, because the land has been extended seaward. On an emerging land area therefore the normal growth of a valley involves its lengthening at its lower end as well as at its upper. The lengthening of a valley, or at least the lengthening of a stream, also takes place at its lower end if the land in which it lies is being extended seaward by deposition.

Structural valleys.—In mountain regions valleys are sometimes formed by the uplift of parallel mountain folds, leaving a depression between (Fig. 58). Drainage will appropriate such a valley so that it becomes in some sense a river valley. But it is not a river valley in the sense in which the term has been used in the preceding pages. It is rather a *structural valley.* In its bottom a river valley may be developed (*a*, Fig. 58).

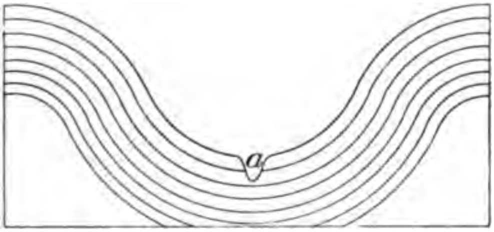

FIG. 58.—Structural valley with a river valley developing its bottom.

The foregoing illustrations by no means exhaust the list of conditions under which valleys develop, but they suffice for the present.

The courses of valleys.—River valleys are rarely straight. To understand why they are crooked it is only necessary to understand the methods by which they grow. In so far as a river valley is a gully grown big, that is, in so far as its length is the result of head erosion, its course was determined by the course of the antecedent gully. If in the case shown in Fig. 59 the slope of the surface above the head of the gully is uniform, its material homogeneous, and the rainfall everywhere equal, more water will come into the gully from the direction *a* than from any other. In this case there would be more wear in the direct line of its extension than elsewhere, and the head would advance in a straight line. But if there be inequalities of slope about the head of a gully at any stage of its development more water may come in

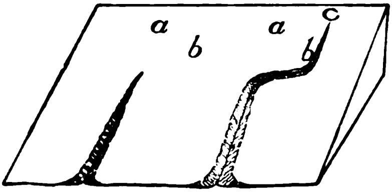

FIG. 59. FIG. 60.

Figures to show why the head of a gully (and therefore a valley) departs from a direct course.

from some direction other than that in the direct line of its extension. In Fig. 59, for example, more water may enter from the direction of *b* than from that of *a*. Since most wear is likely to be affected along the line of greatest inflow, the head of the gully will be turned in that direction (Fig. 60). Started in this course it will continue in the new direction so long as erosion in this line is greater than that elsewhere; but whenever the configuration of the surface causes more water to enter the head of the gully from some direction other than that in which it is headed, the line of axial growth is again changed, as toward *c*, Fig. 60. Since new land surfaces are probably more or less undulatory, crookedness should be the rule among valleys developed from gullies by head erosion. Streams and valleys the courses of which are determined by the original slope of the land are said to be *consequent*.

Inequalities of material, leading to unequal rates of erosion, effect the same result, in the absence of inequalities of slope. If at any stage of a valley's development erosion were equal in two directions at its head, and at the same time greater than at points between, two gullies would result (Fig. 61) diverging from the point in question.

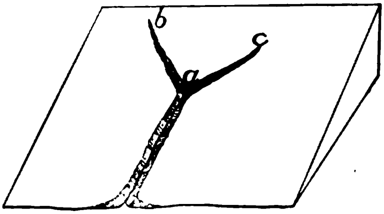

FIG. 61.—Diagram illustrating the development of two equal gullies from the head of one.

In the case of a valley developed by overflow from a lake its course is determined by the lowest line of flow to which the water has access. If this line be straight the valley will be straight; if it is crooked, as it generally is, the valley is crooked also.

The development of tributaries.—Thus far valleys leading immediately to the sea have been considered, and no account taken of tributaries. As a matter of fact most considerable valleys have numerous tributaries. It is now in order to inquire into their mode of development.

So soon as a gully is started, the water flowing into it from either side wears back the slopes. The least inequality of slope, or the least variation in the character of the material, is sufficient to make the lateral erosion unequal at different points, and unequal erosion in the slopes results in the development of tributary gullies. The oldest tributaries may be nearly as old as the main which they join, and from which they developed, for the possibilities of unequal side erosion exist as soon as a gully is opened. While the main gully is developing into a ravine, and the ravine into a valley, the tributary gullies are likewise developing into maturer stages. Tributary to a young valley, therefore, there may be gullies near its head, ravines in its middle course, and small valleys along its oldest portion. It is not to be understood, however, that the oldest tributaries are necessarily the largest, for because of more favorable conditions for growth the younger tributaries often outstrip the older.

The position of tributaries with reference to their mains is worthy of note. The water flowing down a slope follows the line of steepest descent. A gully is usually wider at its lower end, and narrower at its upper. Wherever this is true the line of steepest descent down its side is not a line perpendicular to its axis, but a line slightly oblique to it (*ef*, Fig. 62), and oblique in such a direction that it meets the axis with an obtuse angle below and an acute angle above. It is in the direction corresponding to this line that tributary gullies tend to develop. Thus at the inception of its history a tributary gully is likely to join its main with an angle slightly acute on the up-stream side. If the tributary did not begin until after its main was farther advanced this tendency would be less and less pronounced. Inequalities of material or slope would often counteract this tendency, which, at best, would cause the courses of tributaries to depart but little from perpendicularity to their mains.

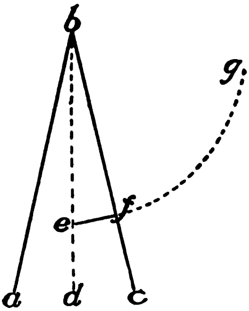

FIG. 62. — Diagram to illustrate the oblique position of a tributary gully at its inception, and its later normal change of direction.

After the head of a tributary has worked back from the immediate slope of its main every condition which determines the course of a gully is likely to affect it, and it is by no means certain that it will continue to lengthen in the direction in which it started. Since the general slope of the surface into which the tributary works is likely to be seaward, more water is likely to enter from the landward than from the seaward side of its head, so that, except where there are notable irregularities of slope, its tendency will be to turn more and more toward the direction of its main (*efg*, Fig. 62).

In depth the tributary is always limited by its main. The principles which determine the length and width of a main valley determine also the length and width of a tributary (see p. 67 et seq.).

A CYCLE OF EROSION. ITS STAGES.

From what has preceded it is clear that the topography of a region undergoing erosion will change greatly from time to time. The first effect of erosion is to roughen the surface by cutting out valleys, leaving ridges and hills. The final effect is to make it smooth again by cutting the ridges and hills down to the level of the valleys.

The base-level of erosion has already been defined; but the mode of its development may now be illustrated in the light of the preceding discussion. Suppose a land surface affected by a series of parallel young valleys without tributaries (Fig. 63). Between them there is a series of upland plateaus. The profile of the surface between two adjacent valleys is represented in section by the uppermost line in Fig. 64. As the valleys are widened from 1–1 and 1′–1′, to 2–2 and 2′–2′, the intervening plateau is correspondingly narrowed. When the valleys have attained the form represented by 3–3 and 3′–3′, the intervening upland has been narrowed to a ridge, *a*, and the valley flats have become wide. With continued erosion the ridge will be lowered (to *b* and below), and

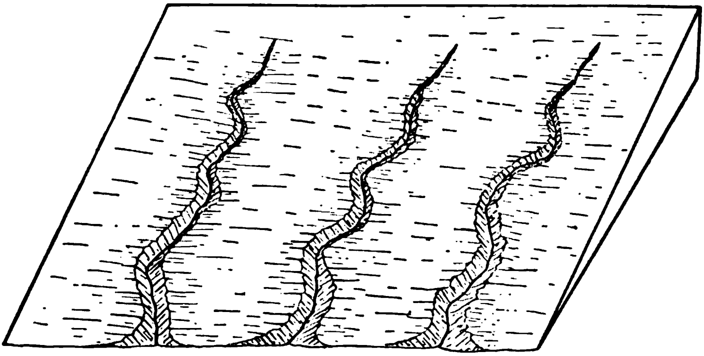

Fig. 63.—Diagram showing three parallel valleys in a land surface.

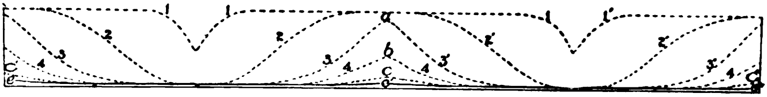

Fig. 64.—Diagram to illustrate the lowering of the surface by valley erosion. The successive cross profiles of the valleys are represented by the lines 1–1, 1–1′, 2–2, 2–2′, etc.

in time the surface will approach a plain. In this condition it is known as a *peneplain* (an "almost-plain"). Finally, when running water has done its utmost, the ridges will be essentially obliterated and a base-leveled plain (e, e', e'') results. The figure expresses the fact that the base-level develops laterally from the axis of the valley. It also develops headward from the seaward end of the valley. Similarly, taking into account all the valleys which affect it, the seaward margin of a base-leveled plain is developed first, and thence it extends itself inland.

Tributaries are tolerably sure to develop along each main valley. The heads of the tributaries work back across the uplands between the main valleys, dissecting them into secondary ridges (Fig. 65). Tributaries will develop on the tributaries, and these tertiary valleys dissect the secondary ridges into those of a lower order. This process of tributary development goes on until drainage lines of the fourth, fifth, sixth, and higher orders are formed (Fig. 66). Since the process of valley development under such circumstances is also the process of ridge dissection, a stage is presently reached where the ridges are cut into such short sections that they cease to be ridges, and become hills instead. Even then the processes of erosion do not stop, for the rain-water falling on the hills washes the loose material from their surfaces, and starts it on its seaward journey. Thus the "everlasting hills" themselves are lowered, and, given time enough, will be carried to the sea. Under these conditions, as under those already discussed,

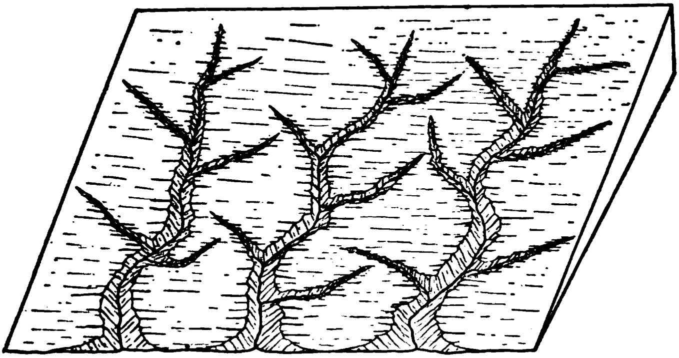

Fig. 65.—Diagram showing the dissection of the upland shown in Fig. 64 by tributary valleys.

the final result of stream erosion is the reduction of the land to base-level. The base-leveled surface, as before, would not be absolutely flat. The area reduced by each stream will have a slight gradient downstream, and from each lateral divide toward the axis of the valley. The crests of the scarcely perceptible elevations which remain will be in the position of the former divides, and these will be highest where most distant from the sea by the course which this part of the drainage took. Even the insensible divides between streams flowing in a common direction may disappear, for when valleys have reached their limits in depth, their streams do not cease to cut laterally. Meandering in their flat-bottomed valleys, they often reach and undercut the divides (Pl. VII), whether they be high or low. By *lateral planation*, therefore, the divides between streams may be entirely eaten away.

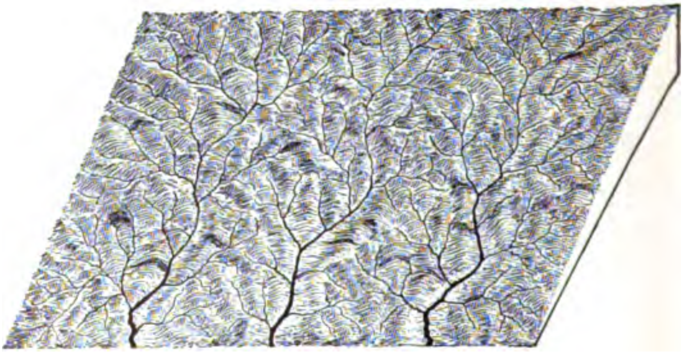

FIG. 66.—Diagram showing tributaries of several orders developed from the conditions sketched in the text.

It has now been seen that by whatever method erosion by running water proceeds, whether there be many valleys, or few or none, the final result of subaërial erosion must be the production of a base-level. It has also been seen that the base-level is first developed at the lower ends of the main streams, and that it extends itself systematically up the main valleys and up all tributaries. The time involved in the reduction of a land area to base-level is a *cycle of erosion*.

It will have been evident from the preceding pages that the terms "grade," "graded plain," and "base-level" and "base-leveled plain," are somewhat variously, and therefore somewhat confusingly, used. "Grade is a condition of essential balance between corrasion and deposition."[1] A graded valley is one in which deposition and corrasion are, in the vertical sense, balanced. Its angle of slope is most variable, and is dependent on the capacity of the stream for work, and on the work it has to do. A weak river must have a higher gradient than a strong one; a stream with much sediment must have a higher gradient than one with little, and a stream with a load of coarse material must have a higher gradient than one with a load of fine. Thus the graded valley of the lower Mississippi has an inappreciable angle of slope, but the graded valleys of many of its tributaries have slopes of hundreds of feet per mile. Since both the size of the stream and the amount and coarseness of its load at a given place vary from time to time, it is clear that the inclination of a graded valley must vary also, and further, that it must be in process of continual readjustment. With the changing conditions of advancing years the slope of a graded valley normally decreases The same principles apply to graded surfaces outside of valleys.

In the continual readjustment of grades incident to a river's normal history the land is brought nearer and nearer to sea-level without ceasing to be at grade. When the inclination of a graded surface becomes so low that it is sensibly flat, the surface may be said to be at *base-level,* although this does not mean that the surface can never be degraded further. If the term be used in this way, it is clear that there is no sharp line of distinction between a graded surface and a base-leveled surface, and as the terms are now commonly applied no such distinction exists.

If the term base-level were made synonymous with sea-level, as has been proposed,[2] the term might as well be discarded, for sea-level could always be used in its stead. Furthermore, streams often erode below sea-level. The bottom of the channel of the Mississippi is below sea-level for some 400 miles above its debouchure, and locally (Fort Jackson) it is nearly 250 feet below. This deep channel is the result of the erosive activity of the stream, not of subsidence. Again, the sea-level is itself inconstant. The extent of its changes cannot now be measured, but they have probably been more considerable in the course of geological history than has been commonly recognized. It is true that they take place slowly, as far as known, but it is also true that the duration of an erosion cycle is sufficiently long for even very slow changes to reach great magnitude. The sea-level, therefore, can hardly be accepted as the absolute base-level, unless (1) the absolute base-level is a variable, and unless (2) the absolute base-level be a surface below which rivers may cut to the extent of at least 250 feet.

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[1] Davis. Jour. of Geol., Vol. X, p. 87.

[2] *Ibid.*, p. 77 et seq.

The ocean may be looked upon as a barrier which in a general way limits the down-cutting of running water; for only very large streams cut much below its level. Other barriers, such as lakes, and the outcrops of hard rock in a stream's bed, have a comparable, though more temporary, effect on the development of valley plains above. Plains thus developed have been called *temporary base-levels.* They differ from other graded plains in being controlled primarily by a barrier below, rather than by conditions which exist above.

Since river valleys have a beginning and pass through various stages of development before the country they drain is base-leveled, it is important to recognize their various stages of advancement. Nor is this difficult. An old valley and a young one have different characteristics, and the one would no more be mistaken for the other by those who have learned to interpret them, than the face of an aged man would be mistaken for that of a child.

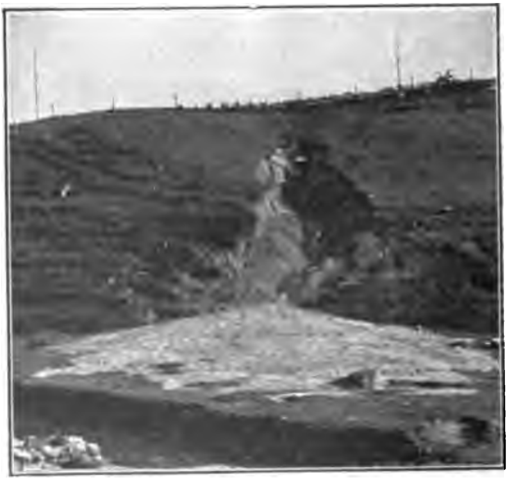

FIG. 67.—A gully developed by a single shower. (Blackwelder.)

The cycle begins with the beginning of valley development, and at that stage drainage is in its *infancy.* The type of the infant valley is the gully or ravine (Figs. 67 and 68). It has steep slopes and a narrow bottom. Fig. 1 of Plate IV represents similar, or rather older, ravines in contour (shore of Lake Michigan, just north of Chicago). With age, the valley widens, lengthens, and deepens, and passes from infancy to *youth*. In this stage also the valleys are relatively narrow, and the divides between them broad. They may be

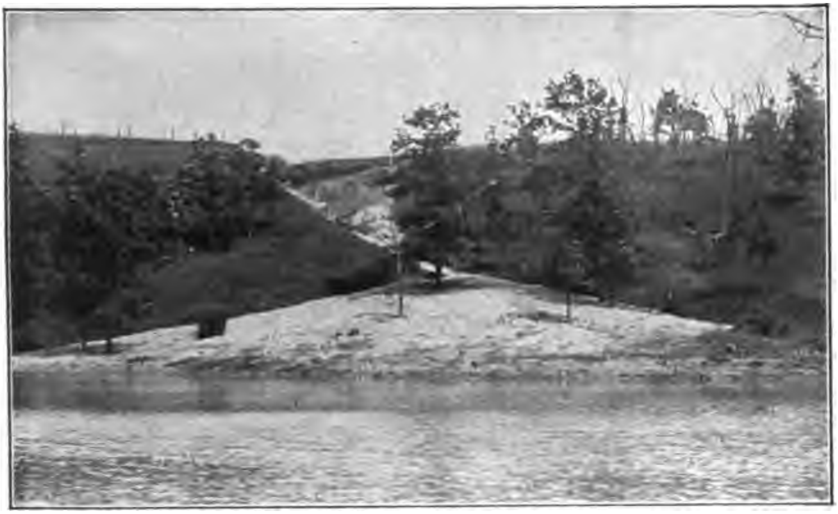

FIG. 68.—A gully somewhat older than that shown in Fig. 67. (Alden.)

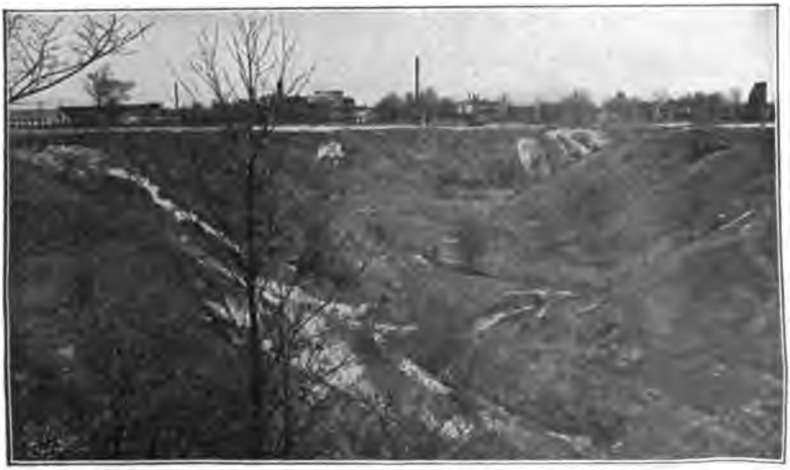

FIG. 69.—A young valley in a region of slight relief.

deep or shallow, according to the height of the land in which they are cut, and the fall of the water flowing through them; but in any case the streams flowing through them have done but a small part of the work they are to do before the country they drain is base-leveled. Figs. 69 and 70, respectively, represent youthful valleys in regions of moderate and great relief. Fig. 2, Plate IV, shows a youthful valley in a region of slight relief (near Casselton, N. D., lat. 46° 40′, long. 97° 25′). The uppermost line in Fig. 64 likewise represents topographic youth, as shown in cross-profile.

Not only are narrow valleys said to be young, but the territory affected by them is said to be in its topographic youth, since but a small part of the time necessary to reduce it to base-level has elapsed. An area is in its topographic youth when considerable portions of it are still unaffected by valleys. Thus the areas (as a whole), as well as the valleys, represented on Plate IV, are in their topographic youth. It is often convenient to recognize various sub-stages, such as early, middle, and late, within the youthful stage of valleys or topographies. The different parts of the areas shown on Plate IV, for example, represent different stages of youth.

Youthful streams, as well as youthful topographies, have their distinctive characteristics. They are usually swift; their cutting is mainly at the bottom rather than at the sides, and their courses are often marked by rapids and falls.

As valleys approach base-level they develop flats. As the valleys and their flats widen, and as their tributaries increase in numbers and size, a stage of erosion is presently reached where but little of the original upland surface remains. The country is largely reduced to slopes. In this condition the drainage and the topography which it has determined are said to be *mature*. Mature topography is shown in contours in the figures of Plate V, and in the northern part of Plate VI, where slopes, rather than upland or valley flats, predominate. Fig. 1 of Plate V represents an area in southeastern Kentucky (lat. 37° 12′, long. 83° 10′); Fig. 2, an area in western Virginia. Plate VI represents an area in southern California, somewhat west of San Bernardino. The three areas are alike in representing mature drainage, though not of equal stages of advancement. The striking differences of topography of the three areas are the result of differences in rock structure and altitude, and will be considered later. Mature topography is also

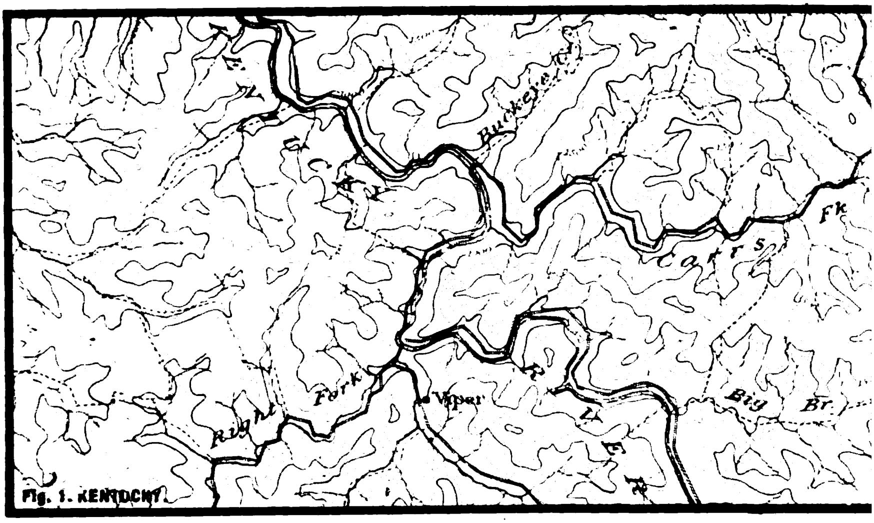

Fig. 1. KENTUCKY.

U. S. Geol. Surv.

Scale, 2+ miles per inch.

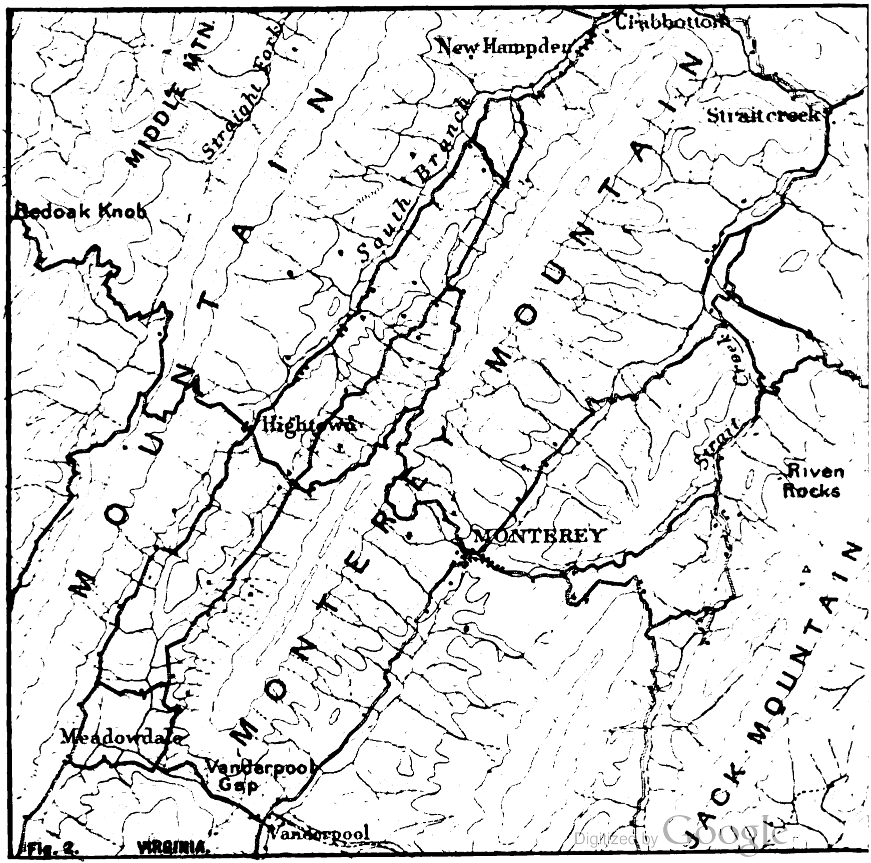

Fig. 2. VIRGINIA.

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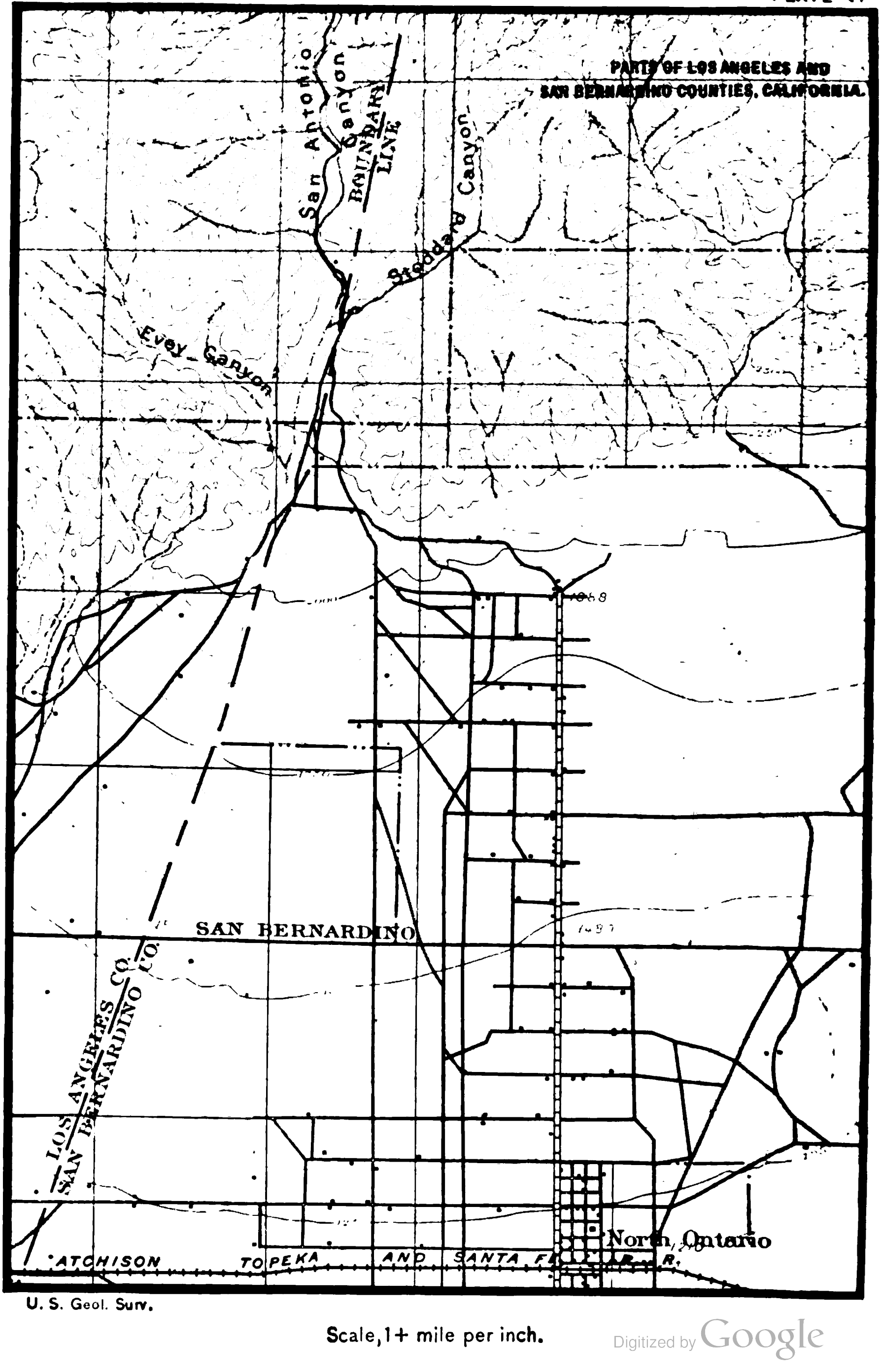

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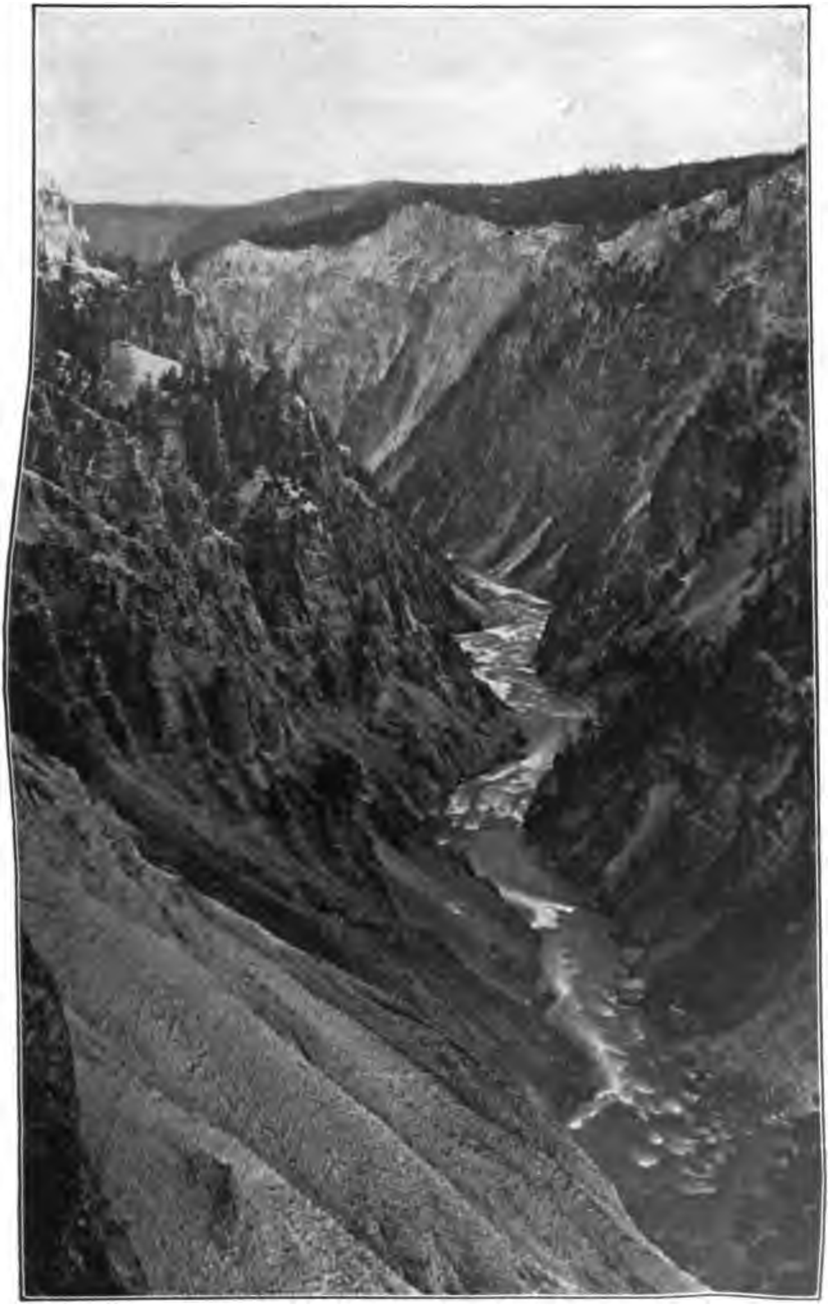

Fig. 70.—The valley (canyon) of the Yellowstone. A young valley in an elevated region. (Page 87.)

shown in Fig. 71, where the relief is moderate, and in Figs. 72 and 73, where it is great. Figs. 72 and 73 illustrate clearly the universal tendency of rivers in regions of notable relief to develop new flats well below the old surface of the region. At the same time that these low-lying flats are developing, tributary drainage is dissecting and roughening the upper surfaces. This process is well shown in Fig. 73. In both Figs. 72 and 73 the summits of the mountains on either side of the valleys appear to have had about the same elevation. The new flat is therefore developed at the expense of the old flat. As will be seen in the sequel, the first flat which a stream develops along its course is usually somewhat above base-level. It is a *graded* flat.

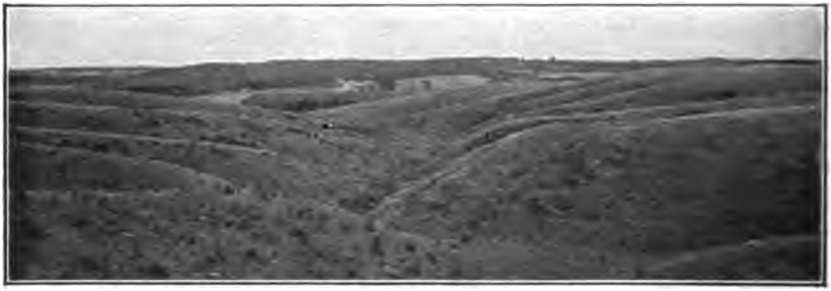

Fig 71.—Mature erosion topography in a region of slight relief, Iowa. (Calvin.)

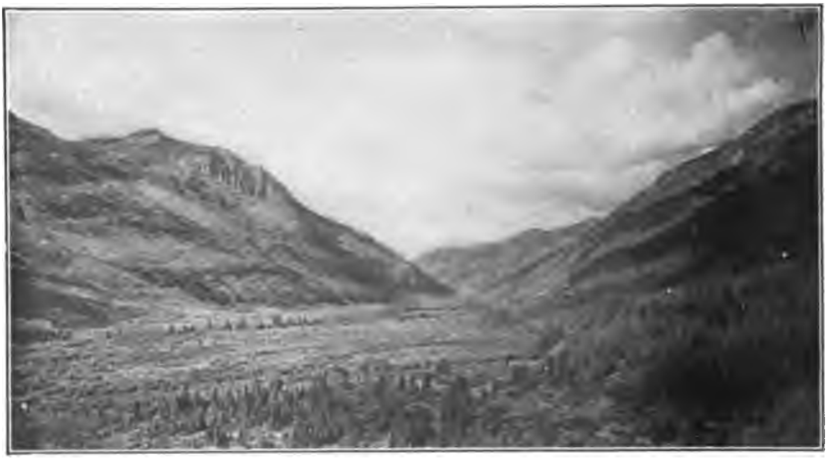

Fig. 72.—Mature erosion in a mountain region. From mouth of Gray Copper Gulch, Silverton, Colo., quadrangle. (Cross, U. S. Geol. Surv.)

The same processes which have made young valleys mature will in time work further changes. When the gradients of the valleys have become low and their bottoms wide, and when the intervening ridges and hills have become narrow and small, the drainage and the drainage topography have reached *old age*, and the streams are in a condition of senility. This is illustrated by Fig. 1, Plate VII (central Kansas), and in section by the third and lower lines in Fig. 64. Topographic old age sometimes has a different expression; this is shown in Fig. 74, where most of the surface has been brought low. The elevations which rise above the general plain are small in area, but have abrupt slopes. This phase of old-age topography is usually the result of the unequal resistance of the rock degraded. The effects of unequal rock-resistance will be considered later.

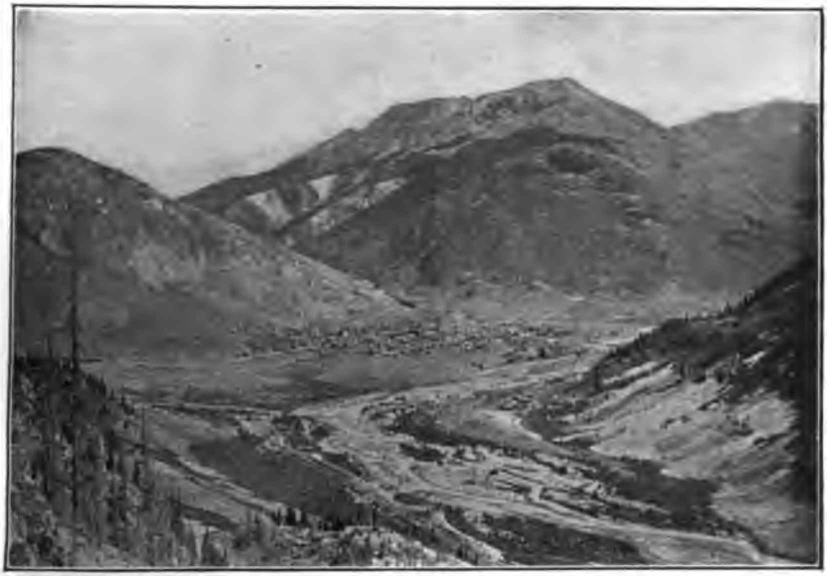

FIG. 73.—Mature erosion in a mountain region. Silverton, Colo. (Cross, U. S. Geol. Surv.)

The marks of old streams are as characteristic as those of young ones. They have low gradients and are sluggish. Instead of lowering their channels steadily they cut them down in flood, and fill them up when their currents are not swollen. They meander widely in their flat-bottomed valleys (Fig. 1, Pl. VII, Central Kansas) and their erosion, except in time of flood, is largely lateral.

If the processes of degradation were to continue until the land surface was brought to sea-level, and this might be done by solution though not by mechanical erosion of running water, the rivers would no longer flow, and the drainage system would have reached the end of its history—*death.*

Not only do valleys normally pass from birth to youth, from youth to maturity, and from maturity to old age, but a single river system may show these various stages of development in its various parts. Thus in the area shown in Fig. 2, Plate VII (north central Kansas), there is a tract (extreme southwest) where the erosion history is scarcely begun. The zone of land a little farther northeast, and just reached by the heads of the valleys (same figure), is in its youth. The well-drained and uneven tract southwest of the flat of the Solomon River is in maturity, while the flat of the main valley has the general characteristics of old age.

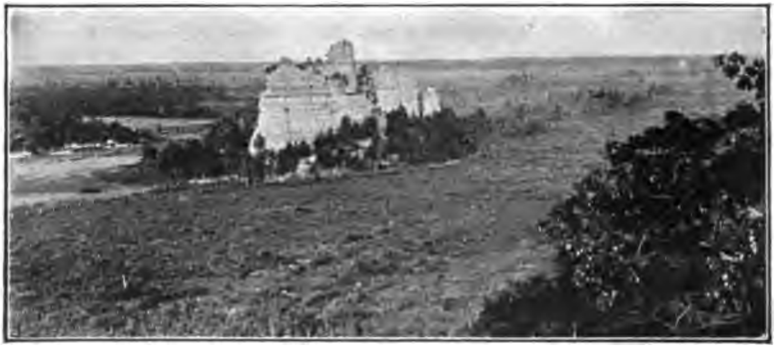

FIG. 74.—A peneplained surface where the elevations are small but steep-sided. Near Camp Douglas, Wis. (Atwood.)

The age of valleys in terms of erosion is also expressed more or less perfectly by their cross-sections. The line 1-1 (and 1′-1′) of Fig. 64 represents in cross-section a narrow V-shaped valley. Such a section is always indicative of youth. The stream which developed it cut chiefly at its bottom, not at its sides. It was therefore rapid, and rapid streams are young. The line 2–2, (2′-2′) (Fig. 64) shows the same valley at a later and maturer stage when downward cutting has nearly ceased. The widening of the valley by slope wash has

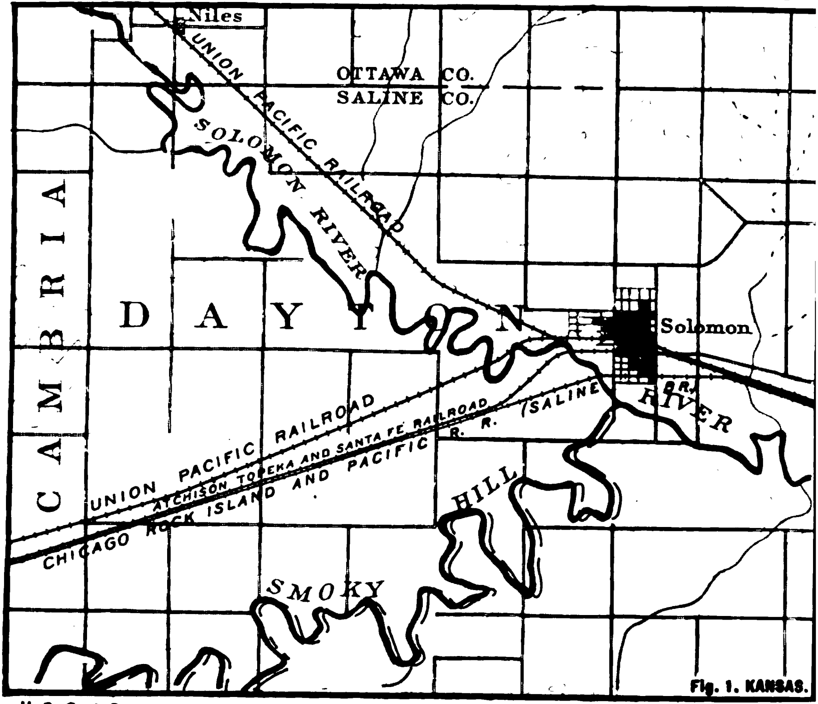

Fig. 1. KANSAS.

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Scale, 2+ miles per inch.

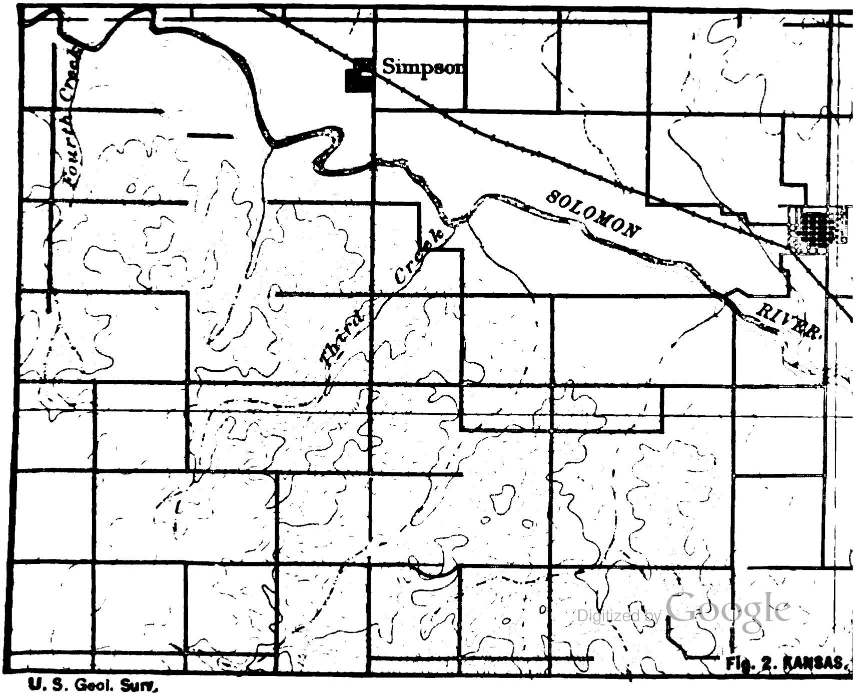

Fig. 2. KANSAS.

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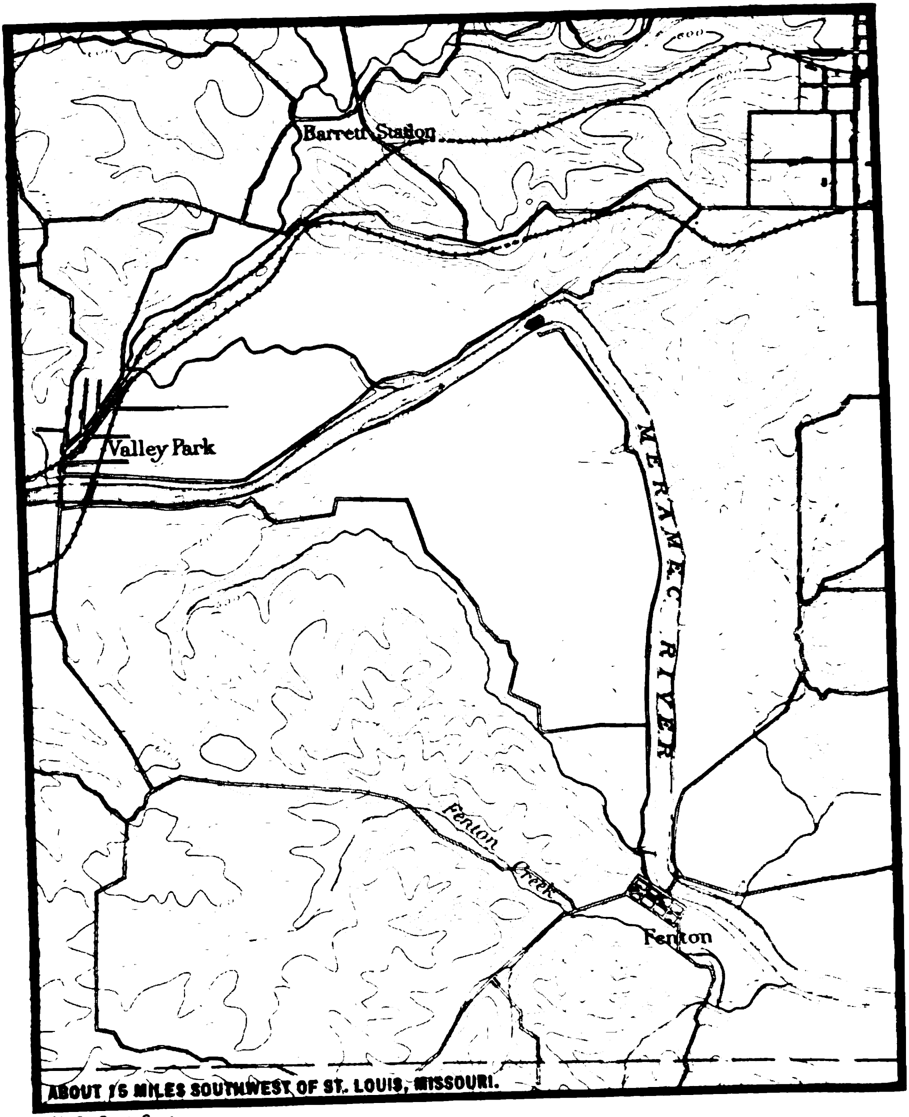

ABOUT 15 MILES SOUTHWEST OF ST. LOUIS, MISSOURI.

U. S. Geol. Surv.

Scale, 1+ mile per inch.

become relatively more important than before, and the stream has so far lost velocity as the result of diminished gradient as to be unable to carry away all the detritus washed down from the sides. As a result of deposition at the bases of the side slopes, a concave curve has been developed. Up the valley from the point where such a section as is represented by 2–2 occurs, the valley may still have a section similar to that represented by 1–1.

Still later stages of development are represented by the cross-sections 3–3 and 4–4. Not only has the valley become larger, but the stream has deposited detritus (not shown in the figure) in the bottom of its valley, developing an alluvial flat. On this flat the stream meanders, and the valley may be widened by the under-cutting of the bluffs wherever the stream in its wanderings reaches them (Pl. VIII, near St. Louis). A valley might possess the characteristics shown by the cross-sections 3–3, 2–2, and 1–1, Fig. 64, in its lower, middle, and upper courses, respectively.

The preceding discussion, and the illustrations which accompany it, give some idea of the topography which characterizes an area in various stages of its erosion history. Whether the valleys are deep or shallow, and the intervening ridges high or low, depends on the original height of the land and its distance from the sea. The higher the land, and the nearer it is to the sea, the greater the relief developed by erosion. A plateau near the sea may become mountainous in the mature stage of its erosion history, while a plain in the same situation would only become hilly. A plateau in the heart of a continent would have less relief in its maturity than one of equal elevation near the sea, since the grade-plain in the former position is higher than in the latter. Plates IV and IX show youthful topography where the relief is relatively slight, and Plate X shows youthful topography where the relief is great. Similarly, Plates V and VI show mature topography where the relief is great, and Fig. 1, Plate III, shows mature topography where the relief is relatively slight.

Topographic youth, topographic maturity, and topographic old age are also indicated in other ways, and especially by the presence of features which rivers tend to destroy. If, for example, the surface of the land, well above the valley bottoms, is marked by numerous ponds and marshes, it is clear that drainage has not yet progressed beyond its early stages, for, unless the lakes be very deep, valleys working back into the land will find and drain them before topographic maturity has been reached. Their presence is evidence that the region where they occur has not yet been thoroughly dissected by erosion lines, and therefore has not reached maturity. Still other marks of topographic youth, such as rapids, falls, etc., as well as marks of topographic maturity and old age, will be mentioned in the following pages.

GENERAL CHARACTERISTICS OF TOPOGRAPHIES DEVELOPED BY RIVER EROSION.

With the characteristics of river valleys and the methods by which they grow clearly in mind it is easy to say whether rivers have been the chief agents in the development of a given topography. River valleys are distinguished from other depressions on land surfaces by their linear form and, leaving out of consideration the relatively insignificant inequalities in a stream's channel, by the fact that any point in the bottom of a river valley is lower than any other point farther up the stream in the same valley, and higher than any point farther down the stream. The second point might be otherwise stated by saying that every valley excavated by erosion leads to a lower valley, or to the sea, or an inland basin. Streams which dry up, or otherwise disappear as they flow, constitute partial exceptions. If, therefore, the depressions on a land surface are linear, lead to other and deeper valleys, and finally to an inland basin, or the sea, and if the elevations between these valleys are such as might have been left by the excavation of the valleys, it is generally clear that rain and rivers have been the chief factors in the development of the topography. If, on the other hand, a surface is characterized by topographic features which streams cannot develop, such as enclosed depressions, or hills and ridges whose arrangement is independent of drainage lines, other agents besides rain and surface streams have been concerned in its development.

SPECIAL FEATURES RESULTING FROM SPECIAL CONDITIONS OF EROSION.

Many striking topographic and scenic features result from rain and river erosion. Some of them depend primarily on the conditions of erosion, such as climate, altitude, etc., while others depend largely on the structure and resistance of the rock. Between these two classes there is no sharp line of demarkation. Illustrations of two types, dependent largely but by no means wholly on conditions independent of the rock, are cited at this point. Others will be mentioned in other connections.

Bad-land topography.—To a type of topography developed in early maturity in certain high regions where the rock is but slightly, though unequally, resistant, a special name is sometimes given. Such regions are termed *bad lands*. Some idea of bad-land topography is gained from Figs. 75 to 78. Bad-land topography is found in various localities in the West, but especially in western Nebraska and Wyoming, and the western parts of the Dakotas. The formations here are often beds of sandstone or shale, alternating with unindurated beds of clay. Climatic factors are also concerned in the development of bad-land topography. A semi-arid climate, where the precipitation is much

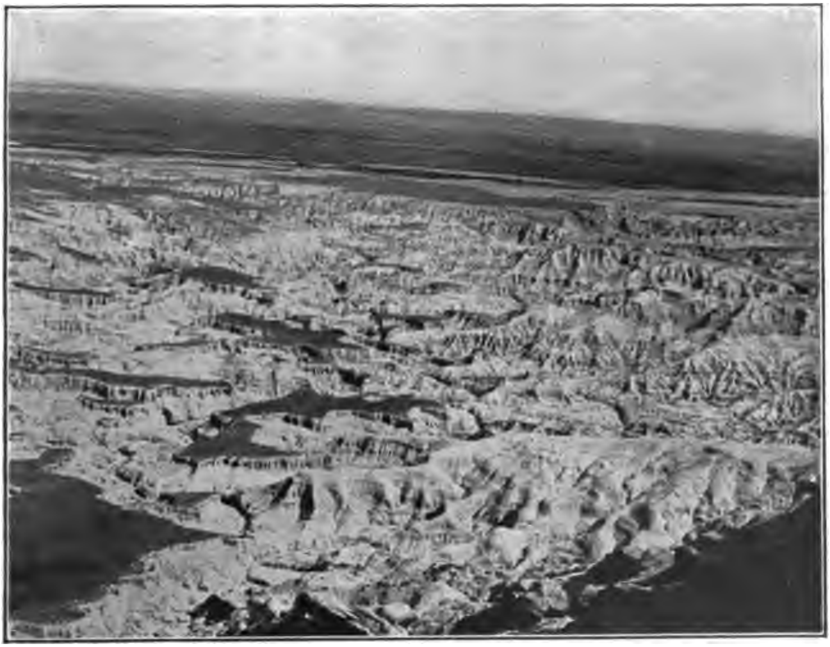

Fig. 75.—Bad-land topography. North of Scott's Bluff, Neb. (Darton, U. S. Geol. Surv.)

concentrated, seems to be most favorable for its development. The bad-land topography is most striking in early maturity.

Special forms of valleys; canyons.—Various conditions influence the size and shape of valleys, especially in the early stage of their development. If the altitude of the land be great, the gradient of the streams at this stage will be high. A high gradient means a swift stream, and a swift stream erodes chiefly at its bottom. High altitudes therefore favor the development of deep valleys. Such valleys will be narrow if the conditions which determine widening are absent or unfavorable. Since slope wash is one of the main factors in the widening of valleys, an arid climate favors the development of narrow valleys, if there be sufficient water to maintain a vigorous stream. Narrowness and steepness of slopes will also be favored if the valley is cut in rock which is capable of standing with steep faces. Thus

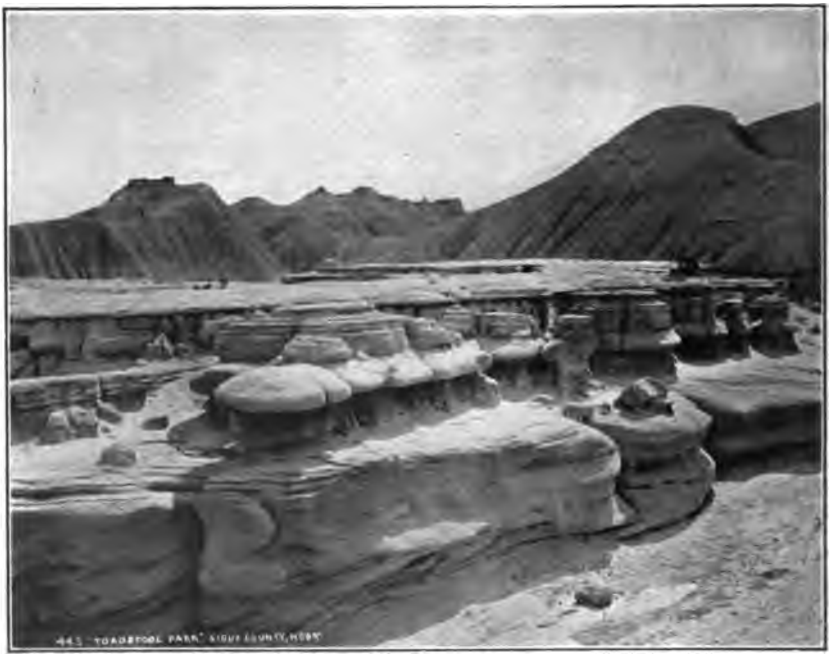

Fig. 76.—Toadstool Park, Sioux Co., Neb. The peculiar topography is the result of erosion working on jointed rocks of unequal hardness in an arid region of considerable elevation where rainfall is unequally distributed. (Darton, U. S. Geol. Surv.)

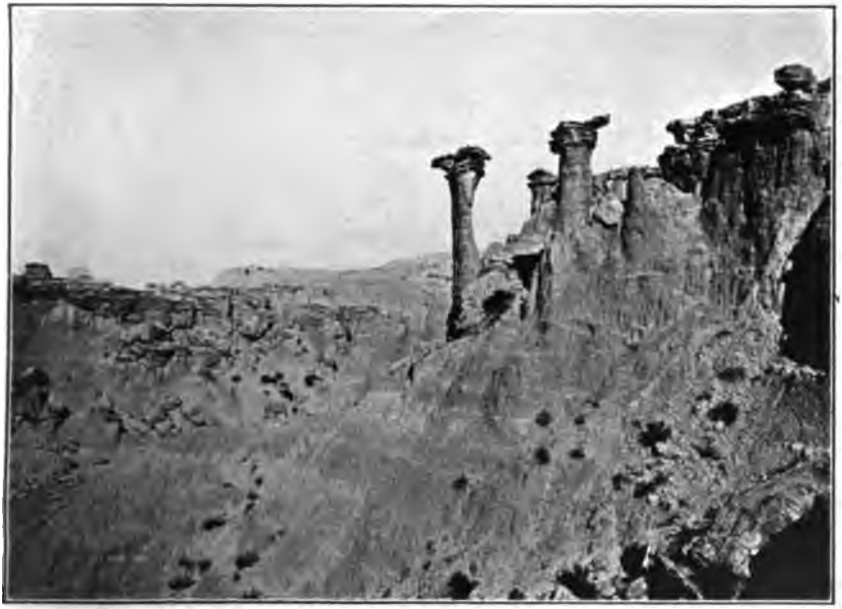

FIG. 77.—Detail of bad-land topography. Head of Indian Draw, Washington Co., S. D. Protoceras sandstone on Oreodon clay. (Darton, U. S. Geol. Surv.)

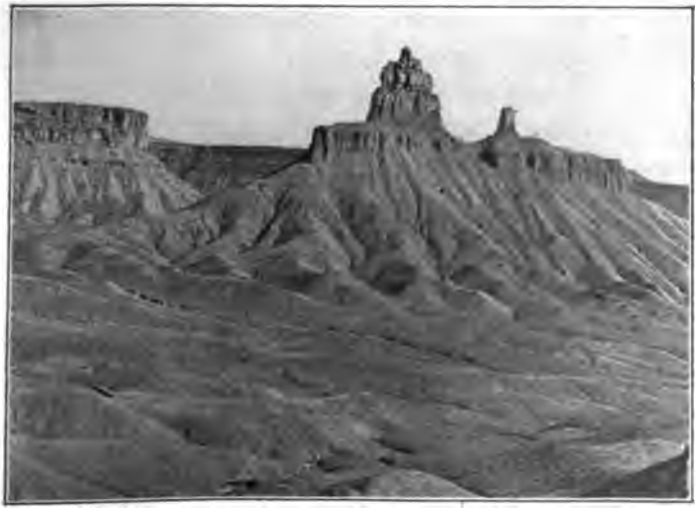

FIG. 78.—Detail of bad-land topography. Southwest foot of Mesa Verde, Colo. (Matthes, U. S. Geol. Surv.)

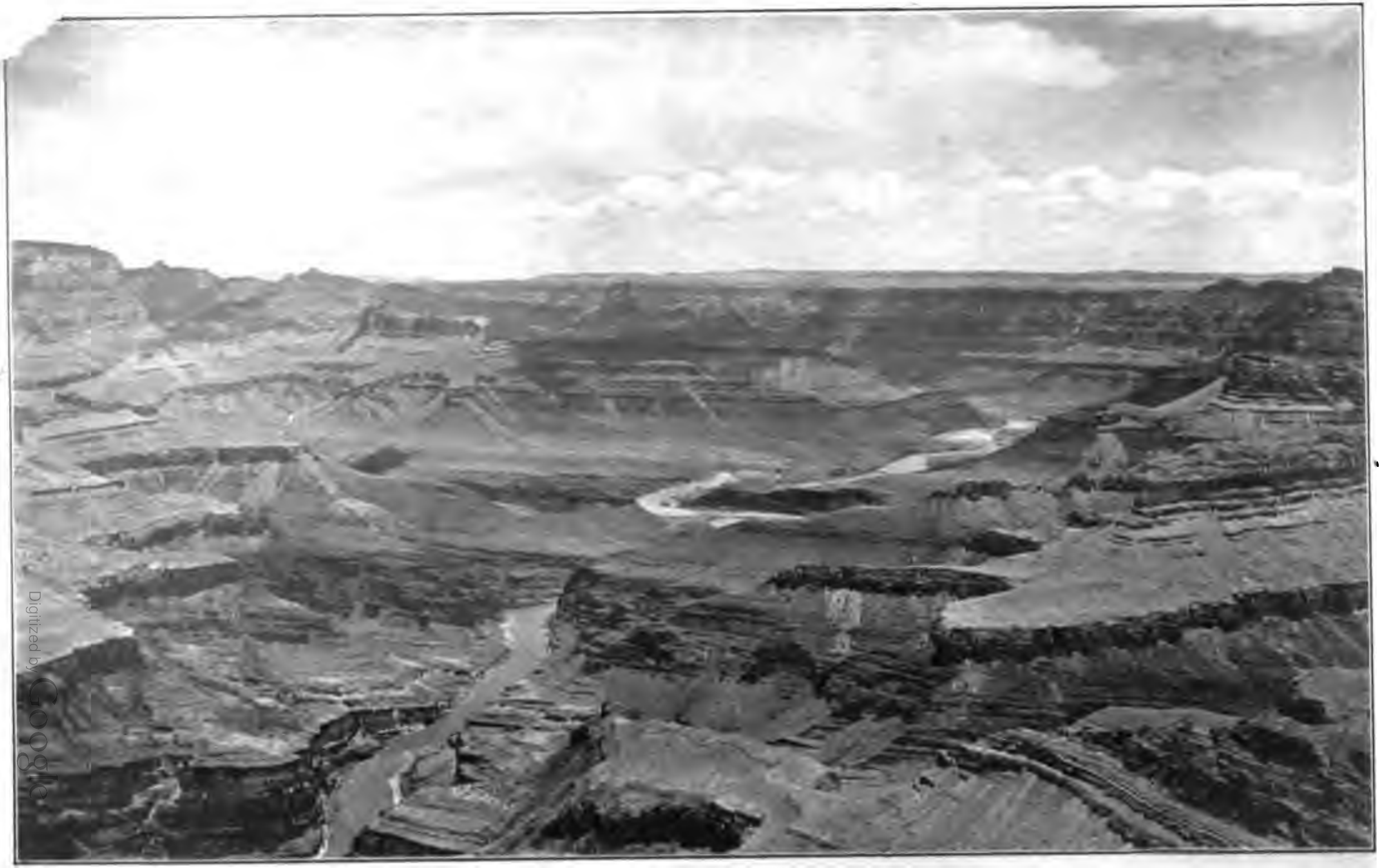

Fig 79.—Grand Canyon of the Colorado. (Peabody.)

(Page 96.)

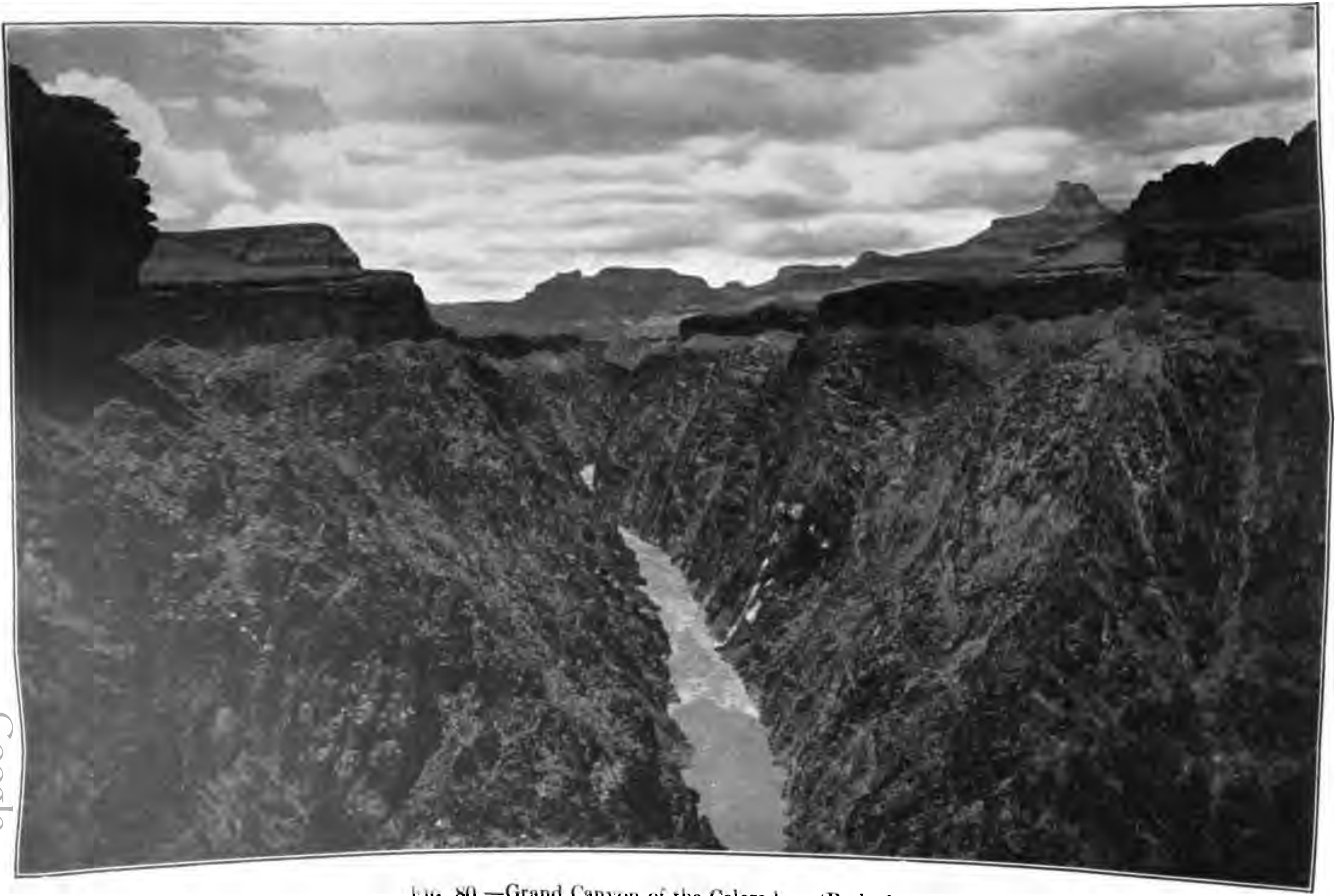

Fig. 80.—Grand Canyon of the Colorado. (Peabody.)

(Page 97.)

a stream may develop a narrow valley in indurated rock where it would not do so in loose gravel, and, other things being equal, it will develop a narrower valley in rock which is horizontally bedded than in rock the beds of which are inclined. Aridity, high altitude, and the proper sort of rock structure therefore favor the development of canyons, and many of the young valleys in the western part of the United States where these conditions prevail, belong to this class.

While all canyons are valleys, most valleys are not canyons. The distinction between a canyon and a valley which is not a canyon is not sharp. The canyon depends for its distinctive character on the relation of depth, width, and angle of slope to one another; but any definition of the depth, width, and angle of slope necessary to constitute a valley a canyon is arbitrary.[1] In popular usage the rule seems to be that if a valley is sufficiently deep, narrow, and steep-sided to be distinctly striking, it is called a canyon in regions where that term is in use. Whether a valley is deep, narrow, and steep-sided enough to be striking clearly depends on the observer. The Colorado Canyon (Figs. 79 and 80) is the greatest canyon known, but it is rarely more than a mile deep, and where its depth approaches this figure it is often eight, ten, or even twelve miles wide from rim to rim. Its width at bottom is little more than the width of the stream; that is, a few hundred feet. Its cross-profile throughout much of its course is therefore not in keeping with the conventional idea of a canyon. With a depth of one mile and a width of eight, the slope, if uniform, would have an angle of less than 15°. Such a valley is represented in Fig. 81. As a matter of fact the slopes of a canyon are not commonly uniform. The slopes represented in Fig. 82 correspond more nearly than those of Fig. 81, to the actual slopes of the Colorado Canyon. The inequalities of slope are occasioned by the inequalities of hardness. It is perhaps needless to say that to an observer on the rim of the canyon the slopes seem several times as steep as those shown in the diagrams.

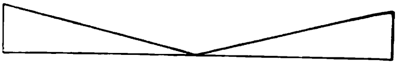

FIG. 81.—Diagram showing the relations of depth and width of a valley, the width of which is eight times the depth.

Like all valleys which are narrow relative to their depth, the Colorado Canyon, great as it is, is a young valley; for it represents but a small part of the work which the stream must do to bring its drainage basin to base-level.

---

[1] In regions where canyons are common, the term is often applied to all valleys.

While aridity and altitude are conditions which favor the development of canyons, as shown by the fact that most canyons are in high and dry regions, they are not indispensable. Niagara River has a canyon below its falls (Pl. IX), and the surrounding region is neither high nor arid. The narrow part of the valley has been developed by the recession of the falls, and is so young that side erosion has not yet widened the valley or lowered its angle of slope to such an extent

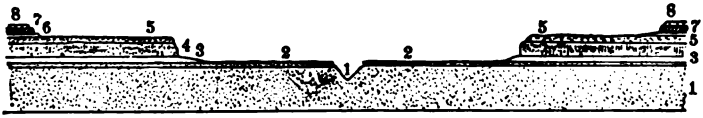

Fig. 82.—Cross-section of the Colorado Canyon. (After Gilbert and Brigham.)

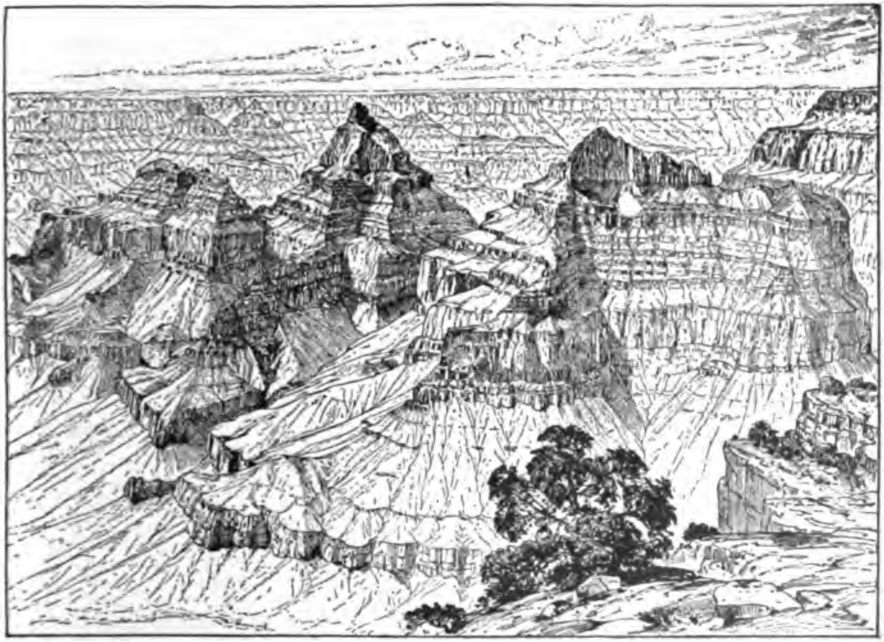

Fig. 83.—Detail of erosion in the Grand Canyon. The inequalities of slope are the result of unequal hardness. The vertical planes which give the architectural effect are the result of joints. (Holmes.)

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as to destroy its canyon character. This canyon is often called a *gorge,* a term frequently applied to small valleys of the canyon type.

Plate X shows portions of the canyons of the Yellowstone and the Colorado rivers respectively. In the first the contour interval is 100 feet, and in the second, 250 feet. The horizontal scale is $\frac{1}{125000}$ (about 2 miles to the inch) in the first, and $\frac{1}{250000}$ in the second. These scales should be borne in mind in interpreting the map.

Falls, rapids, narrows, and other peculiar features, due primarily to inequalities in the hardness of the rock affected by erosion, will be considered later.

THE STRUGGLE FOR EXISTENCE AMONG VALLEYS AND STREAMS.

It is not to be inferred that every gully becomes a valley, nor that every small valley becomes a large one. Among valleys, as among living things, there is a struggle for existence, and fitness determines

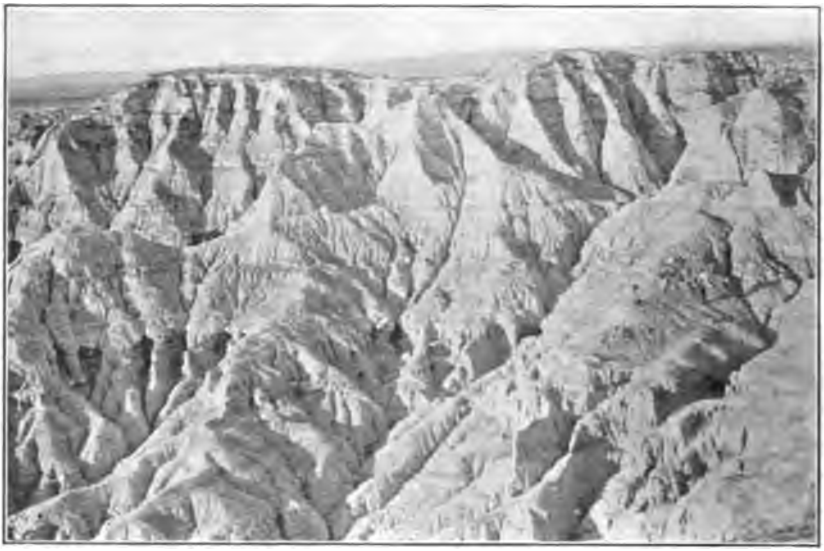
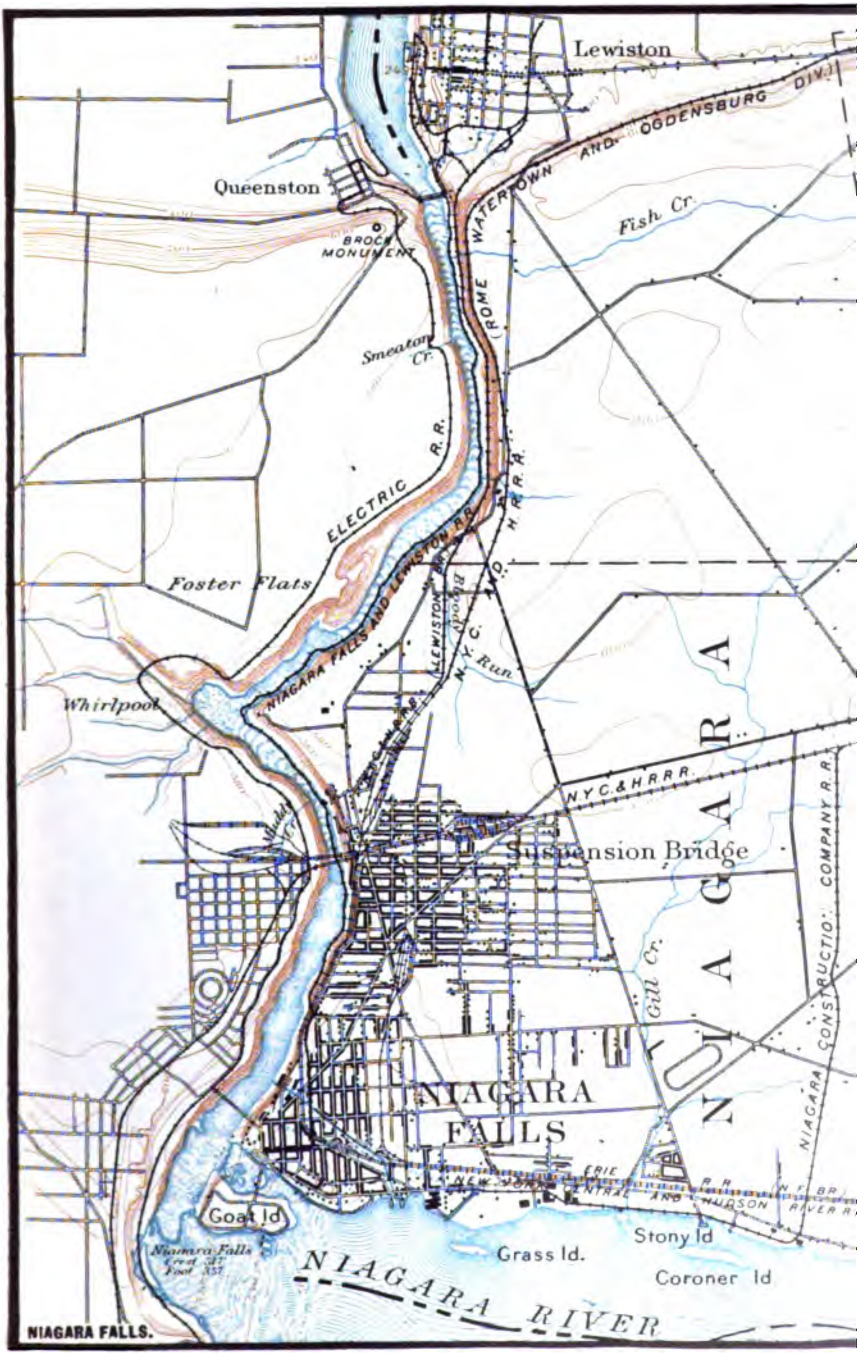


FIG. 84.—A surface illustrating the struggle for existence among gullies. Most of the smaller gullies shown on the slope can have but little growth before being absorbed by their larger neighbors. A type of erosion surface common in the Bad Lands. Scott's Bluff, Neb. (Darton, U. S. Geol. Surv.)

PLATE I



U. S. Geol. Surv.

Scale, 1+ mile per inch.

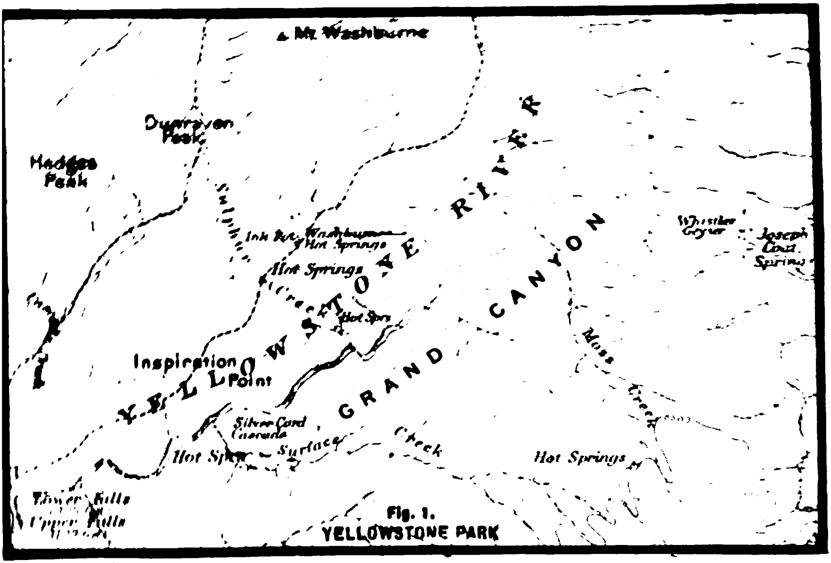


Fig. 1.
YELLOWSTONE PARK

U. S. Geol. Surv.

Scale, 2+ miles per inch.

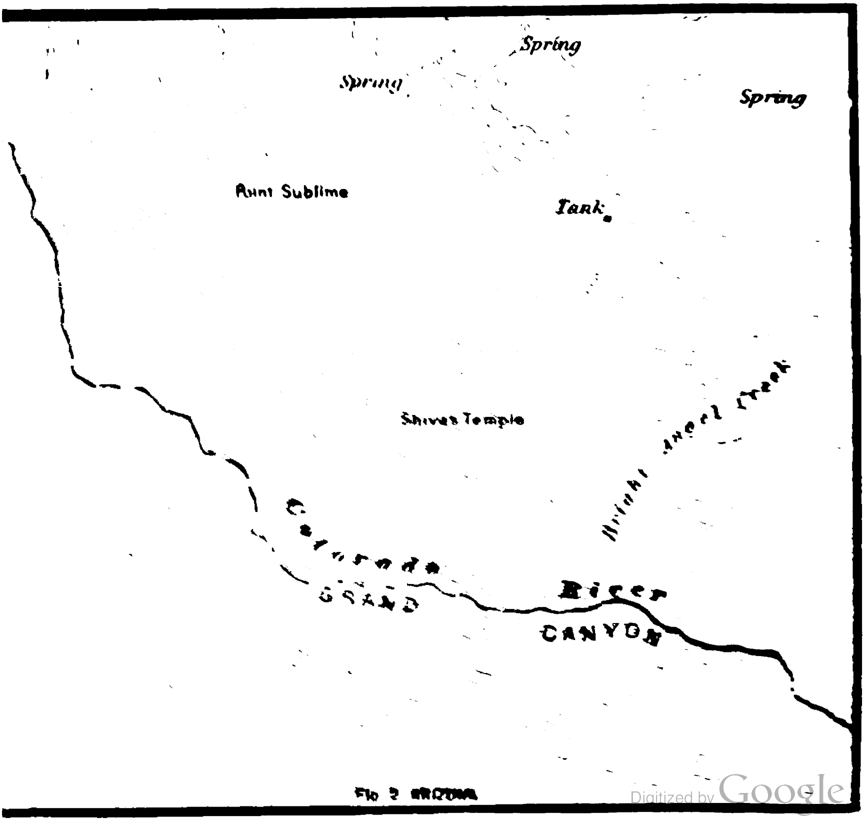


Fig. 2 ARIZONA

U. S. Geol. Surv.

Scale, [illegible] miles per inch.

growth and survival. At an early stage of its erosion history the number of small valleys in a given area is often great, while at a later stage the number is less and the size of the survivors greater.

One phase of the struggle for existence is often well illustrated on a freshly exposed slope of clay. The number of miniature gullies which develop on such a slope, even in a single shower, may be very large (Fig. 84); but the history of many of them is ephemeral. If two adjacent ones are of unequal depth the widening of the deeper narrows and finally eliminates the divide between them, and the two become one (Fig. 85).

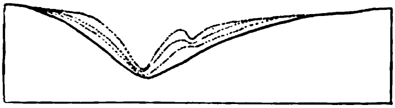
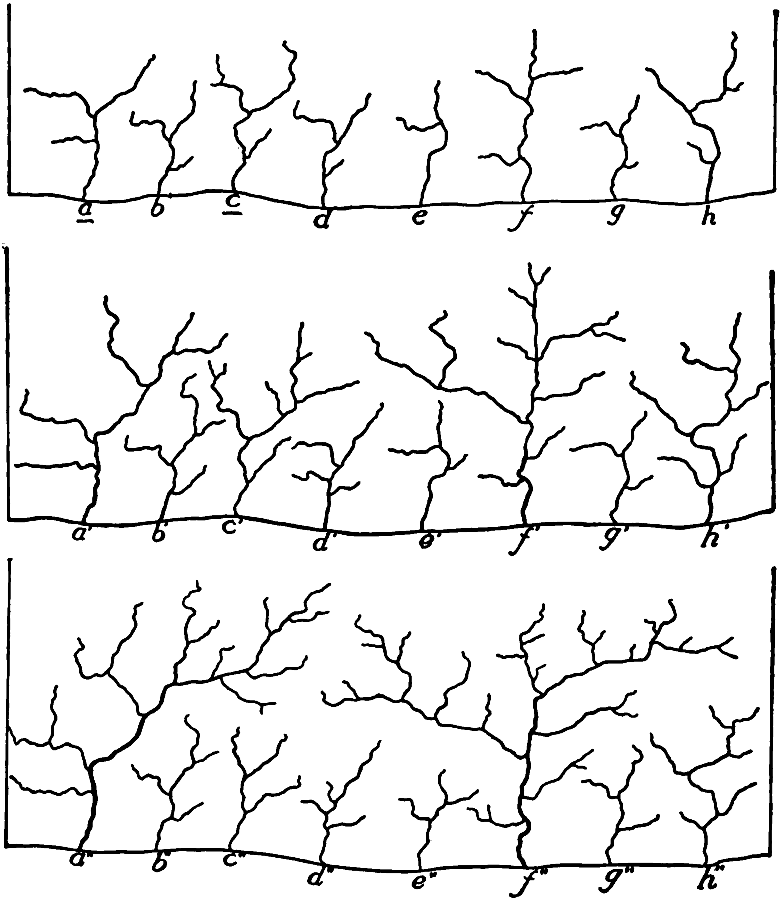


Fig. 85.—Diagram illustrating the absorption of one gully by another by lateral erosion. The successive lines represent successive cross-sections.

Another phase of the struggle for existence is shown in other situations. Examination of a good map of the north shore of Lake Superior or the west shore of Lake Michigan shows a large number of small streams and gullies (Fig. 1, Pl. IV). The valleys are short and narrow, and between and beyond them are considerable areas untouched by erosion. The drainage near the lake is therefore young, and each of the small valleys is growing. This condition of things is perhaps typical of that which has been, is, or will be along the average coast at a certain stage in its erosion history. No equal stretch of coastline where erosion is far advanced can boast of a number of large rivers comparable to that of the many small ones along the coasts mentioned. It therefore seems evident that of these many small streams a few only will attain considerable size.

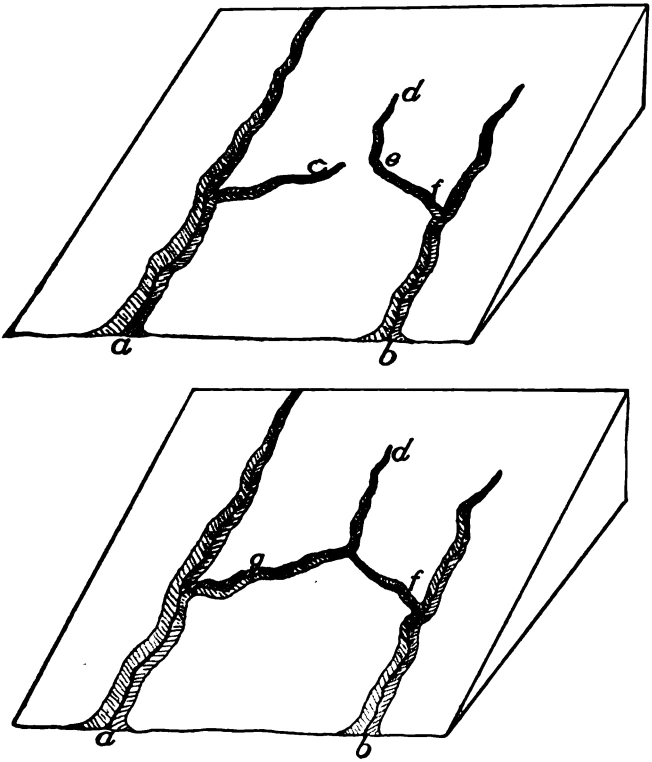
Some of the methods by which the growth of the many is arrested are easily understood. Some of the young valleys on a given coast will work their heads back into the land faster than others because of inequalities of slope and material. This will be true of the tributaries no less than of their mains. If valleys develop in ways other than by head erosion (see p. 73) the chances are also against their equality of growth. If two streams, such as *a* and *c*, Fig. 86, develop faster than the intermediate stream *b*, it is clear that their tributaries may work back into the territory which at the outset drained into *b*, so as to cut off the supply of water from the latter stream (compare *a′b′c′*, Fig. 87). As a result, the growth of *b* will be checked, and ultimately stopped. Similarly other valleys, such as *f*, will get the better of their neighbors, and many of the competitors, as *b*, *d*, *e*, and *g* will soon drop out of the race. Between the stronger streams competition still goes on. If *a'* and *f'* develop faster than *c'* its prospective drainage territory will be preëmpted by its rivals (compare Figs. 87 and 88). Thus as the result of the unequal rate at which valleys are lengthened, the larger number of those which come into existence are arrested in their development. As a result of growth in the manner indicated, the basins of even the large streams remain narrow at their lower ends while they expand above. This is the usual form of a drainage basin the development of which has been normal.



FIGS. 86–88.—**Diagrams to illustrate successive stages in the struggle for existence and dominion among streams.**

Did valleys grow in length only, competition would not destroy the small ones; it would simply limit them. But valleys widen as well as lengthen, and by widening, adjacent valleys may eliminate the divide between them and become one. The elimination of the intervening ridge may be by lateral planation (p. 82), or, if the valleys be of unequal depth, by slope wash (see Fig. 85). By these and other processes many young valleys are dwarfed, and many others are destroyed.

Piracy.—Streams do not always hold the courses which they establish for themselves at the outset. If the valley occupied by the stream *a*, Fig. 89, is deepened more rapidly than the valley occupied by *b*, a tributary from the former, *c*, may work back across the interstream area to *e* and steal the head waters of that stream (Fig. 90). The tributary which does the stealing is known as a *pirate*. Stream *f*



FIGS. 89 and 90.—Diagrams to illustrate piracy.

(Fig. 90) is said to be *beheaded*, and its upper portion, *de*, diverted. The beheaded stream is diminished in volume; or if its total supply of water came in above the point of tapping it would disappear altogether.

The process may not end even here. If after the diversion of *de* the point in the channel to the left is lowered faster than the channel of the beheaded stream *f*, the divide between *dg* and the head of *f* (Fig. 90) will be shifted down the valley of the latter, as shown in Fig. 91. The shifting will go on until the divide reaches a position of stability, that is, until erosion on its opposite sides is equal.

The foregoing case may be called *foreign piracy* because the valleys of different systems are concerned. *Domestic piracy* may also take place, as illustrated in the accompanying diagrams (Figs. 92 and 93). Here a tributary to a crooked river may develop, working back until it taps the main at a higher point, thus straightening the course of the stream. The change takes place only when the highest point in the tributary valley is brought below the surface of the water in the main stream at the point where the tapping takes place. This would be likely to occur only after the main stream had attained a low gradient, for so long as it is deepening its channel notably, the small amount of water flowing through the tributary valley would not be likely to bring it down to the level of the main. In any case the flow of water from the main stream through the new valley would be likely to be started during flood, and at such time the erosion in the new channel

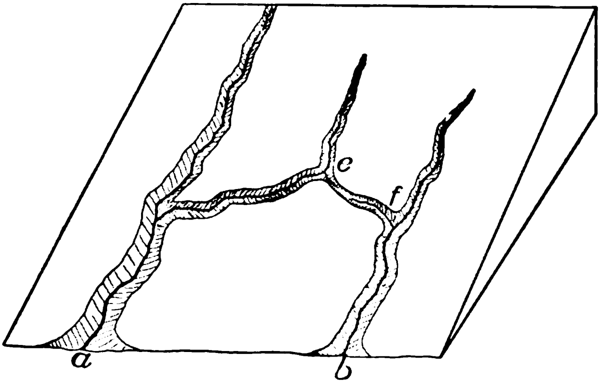
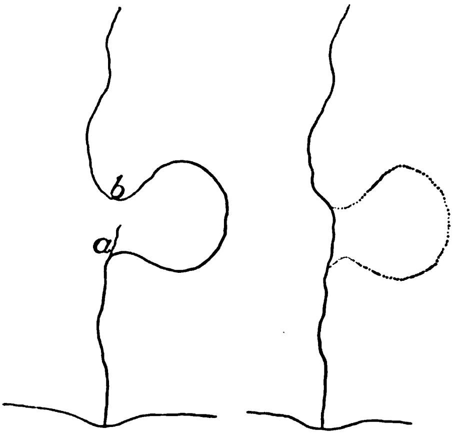


FIG. 91.—Diagram showing the shifting of a divide after piracy.

would be great. The complete and final diversion of the stream through the new channel might be a slow process.

Piracy may occur where the material in which the valleys are cut is homogeneous; but, as will be seen later, heterogeneity of material, by determining unequal rates of erosion, stimulates the piratical proclivities of streams.

An actual case of piracy is shown on Plate XI. North and South Lakes formerly drained westward to the Schoharie Creek, the present head of which is in the extreme northwest corner of the map. The head of Kaaterskill Creek, which had a much higher gradient, worked back and captured the head of the westward-flowing stream, diverting the drainage from North and South Lakes to itself. Schoharie Creek was thus beheaded.



Figs. 92 and 93.—Domestic piracy. The tributary, *a* of Fig. 92, develops headward until it taps the main stream at *b*, giving the result shown in Fig. 93.

Plaatekill Creek, near the south limit of the map, appears to have beheaded the creek flowing west and northwest, similarly diverting its head waters. The Dells, Wis., quadrangle (U. S. Geol. Surv.) affords an illustration of domestic piracy.

RATE OF DEGRADATION.

The amount of mechanical sediment which the Mississippi River carries to the Gulf of Mexico is estimated to represent a rate of degradation for the Mississippi basin of about one foot in 5000 years. But the mechanical sediment carried to the Gulf does not really represent the total degradation of the basin, for the water which sinks beneath the surface is dissolving more or less rock substance, especially lime carbonate. This material is carried to the sea in solution, and does not appear in the sediment on which the above estimate is based. Taking into account the matter dissolved by the water and carried to

the sea in solution, the average rate of degradation for the Mississi
basin is estimated at one foot in 3000 to 4000 years.

It is not to be inferred that this rate is uniform, or even that ero
at any rate whatsoever is taking place in all parts of the basin. S
is not the fact. On the whole the rate of erosion is doubtless grea
toward the margins of the basins where the land is in its topogra
youth or early maturity. It is notably less in the middle cou
of the valleys, and erosion is locally exceeded by deposition a
the lower courses of the Mississippi and some of its main tributa

The average elevation of North America is not accurately kn
but it is probably not far from 2000 feet. If the present rate of de
dation, say one foot in 3500 years, were to continue, it would
something like 7,000,000 years to bring the continent to sea-le
But this rate of degradation could not continue to the end, for as
continent became lower streams would become sluggish and ero
less rapid. Long before the continent reached base-level the rat
degradation, so far as dependent on mechanical erosion, would bec
so slow that the time necessary to bring the continent to sea-l
would be almost inconceivably prolonged. Furthermore, it is q
possible that the land is suffering, or is liable to suffer, uplift, rela
or absolute. If the rate of rise were equal to the rate of degrada
the average height of the continent would of course not be affec

The amount of sediment carried by streams in suspension varies not
according to the stage of the water. During a year when the stream was u
careful study the Mississippi at Carrollton (Miss.) was found to carry $\frac{1}{1}$
its weight of sediment during the high-water stage of June, and $\frac{1}{3381}$ du
the low-water of October, the average for the year being $\frac{1}{1808}$. The ave
of a greater number of records gives about $\frac{1}{1500}$ as the average ratio bet
the weight of the sediment and the weight of the water. This correspond
about $\frac{1}{2900}$ by volume, the average specific gravity being about 1.9.
amount of material carried in the upper part of the water was notably less
that carried at greater depths, but that carried midway between top and bo
was about the same as that carried at the bottom.[1]

The discharge of the Mississippi River is about 19,500,000,000,000 cubic
of water per year, and the sediment it carries in suspension is estimated to w
about 812,500,000,000 pounds. This is equivalent to about 6,714,694,400 c
feet. It is estimated that about 750,000,000 cubic feet of sediment is r
along the bottom, giving a total of 7,468,694,400 cubic feet as the aggre
annual load carried to the Gulf by the river. This would be adequate to c
an area one square mile in extent to the depth of 268 feet per year.

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[1] Humphreys and Abbot. Physics and Hydraulics of the Mississippi River.

vare at Tren- J. .... N. J., 702	Hudso .... C. F. C Public Paper., Am. Healt	Rio Grande del Norte Fort Craig, New Mexico 1873 O. Loew U. S. Geog. Surv. west of 100th M., Vol. III, p. 576	S Hyd m Sept W. Rep Bo 187	e o. 9, Geol. 29	Mohawk Utica, N. Y. C. F. Chandler Johnson's Cyclopedia, Vol. IV	Genesee Rochester, N. Y. C. F. Chandler Johnson's Cyclopedia, Vol. IV
72	.0	.03220			.0036	.0044
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35	.0	.00123			.0069	.00896
21	.0	.03604			.0023	.0024
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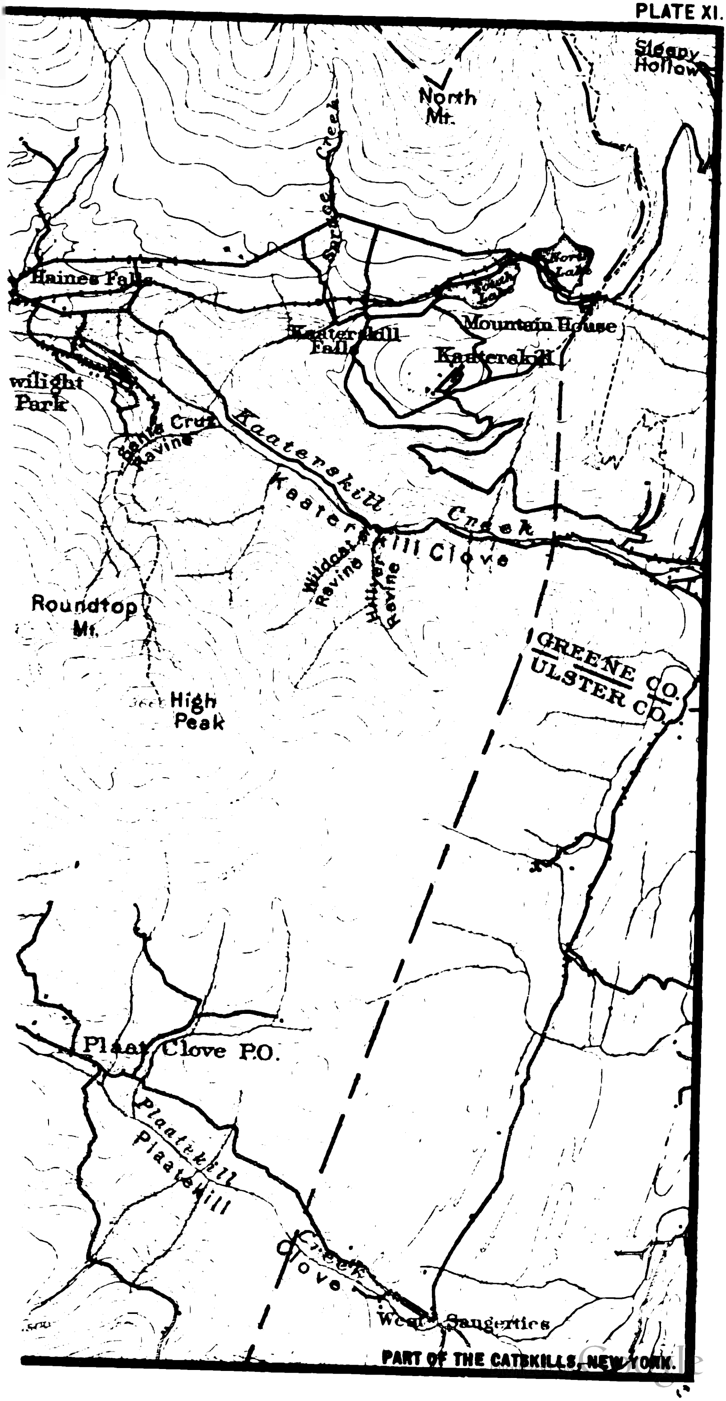
Carbonic acid by difference.



PART OF THE CATSKILLS, NEW YORK.

Scale, 1+ mile per inch.

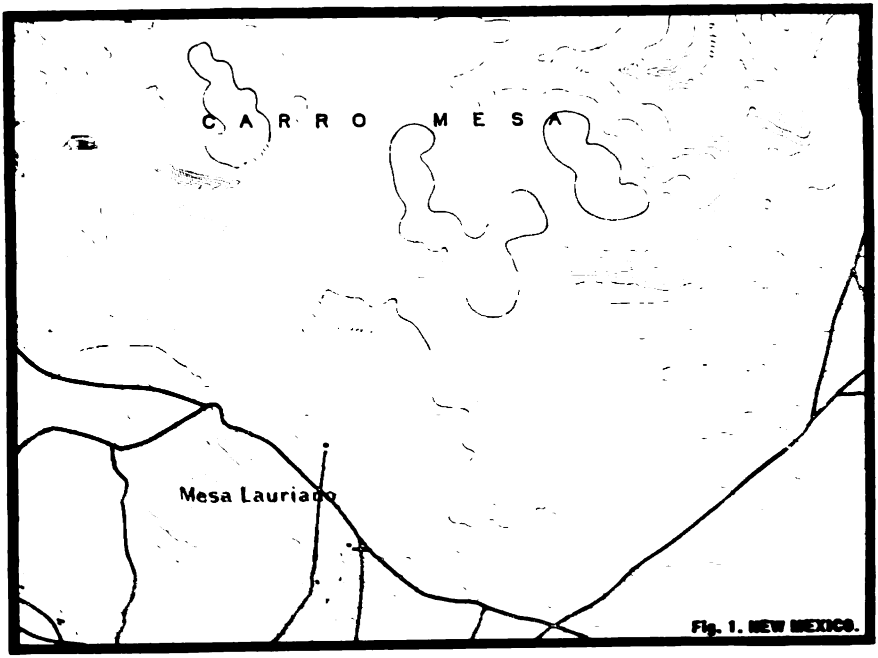


Fig. 1. NEW MEXICO.

U. S. Geol. Surv.

Scale, 2+ miles per inch.

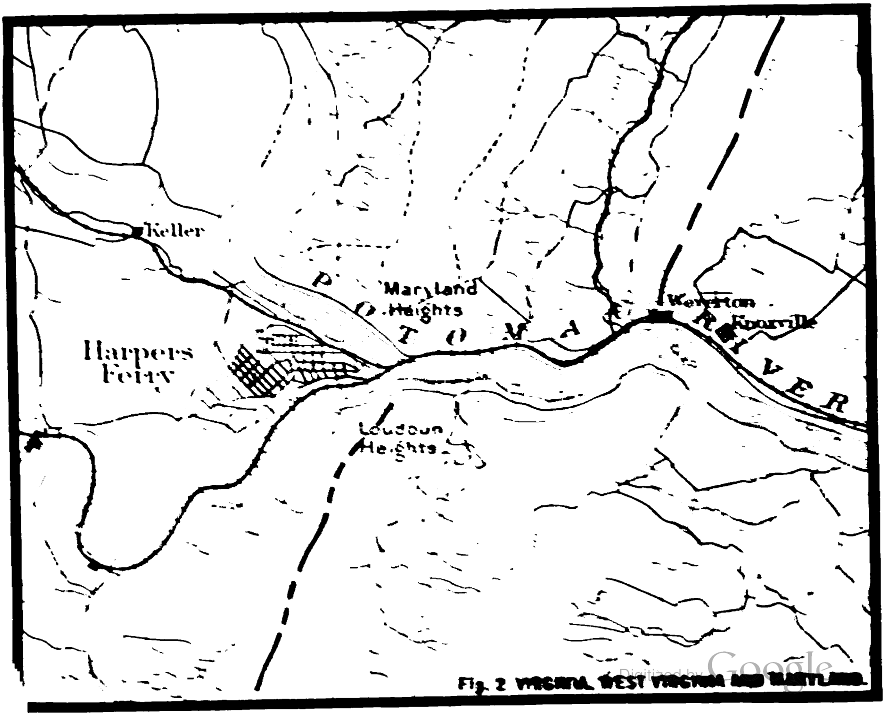


Fig. 2 VIRGINIA, WEST VIRGINIA AND MARYLAND.

U. S. Geol. Surv.

Scale, 2+ miles per inch.

The following table[1] gives the percentage of material carried in suspension by various rivers:

River.	Drainage Areas in Square Miles.	Mean Annual Discharge (in Cubic Feet) per Second.	Total Tons Annually.	Ratio of Sediment to Water by Weight.	Height in Feet of Column of Sediment with a Base of One Square Mile.	Thickness of Sediment in Inches if Spread over Drainage Area.
Potomac	11,043	20,160	5,557,250	1 : 3,575	4.0	.00433
Mississippi	1,244,000	610,000	406,250,000	1 : 1,500	241.4	.00223
Rio Grande	30,000	1,700	3,830,000	1 : 291	2.8	.00116
Uruguay	150,000	150,000	14,782,500	1 : 10,000	10.6	.00085
Rhone	34,800	65,850	36,000,000	1 : 1,775	31.1	.01075
Po	27,100	62,200	67,000,000	1 : 900	59.0	.01139
Danube	320,300	315,200	108,000,000	1 : 2,880	93.2	.00354
Nile	1,100,000	113,000	54,000,000	1 : 2,050	38.8	.00042
Irrawaddy	125,000	475,000	291,430,000	1 : 1,610	209.0	.02005
Mean	334,693	201,468	109,649,972	1 : 2,731	76.65	.00614

The composition of rain-water falling near London, as determined by analysis, was as follows:[2]

Organic carbon	.99	part in 1,000,000 of water.
Organic nitrogen	.22	" " " " "
Ammonia	.50	" " " " "
Nitrogen as nitrates and nitrites	.07	" " " " "
Chlorine	6.30	parts in " " "
Total solids	39.50	" " " " "

A comparison of the composition of rain-water with that of springs and rivers gives some idea of the solvent work of water. From a study of the water of nineteen of the principal rivers of the world Murray has compiled the following table[3] showing the amount of mineral matter in average river water:

MATERIAL IN SOLUTION IN ONE CUBIC MILE OF AVERAGE RIVER WATER.[4]

Constituents.	Tons in a Cubic Mile.
Calcium carbonate ($CaCO_3$)	326,710
Magnesium carbonate ($MgCO_3$)	112,870
Calcium phosphate ($Ca_3P_2O_8$)	2,913
Calcium sulphate ($CaSO_4$)	34,361
Sodium sulphate (Na_2SO_4)	31,805

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[1] Babb. Science, Vol. XXI, p. 343. 1893.

[2] Quoted by Mason. Water-supply, p. 204.

[3] Sot. Geog. Mag., Vol. III, p. 76. 1887.

[4] Acids and bases combined according to the principles indicated by Bunsen.

Potassium sulphate (K_2SO_4)	20,358
Sodium nitrate ($NaNO_3$)	26,800
Sodium chloride ($NaCl$)	16,657
Lithium chloride ($LiCl$)	2,462
Ammonium chloride (NH_4Cl)	1,030
Silica (SiO_2)	74,577
Ferric oxide (Fe_2O_3)	13,006
Alumina (Al_2O_3)	14,315
Manganese oxide (Mn_2O_3)	5,703
Organic matter	79,020
Total dissolved matter	762,587

Murray also estimates that the aggregate amount of water flowing into the sea annually is about 6528 cubic miles, which, on the above basis, would carry about 4,975,000,000 tons of mineral matter in solution.

A large number of analyses of waters of rivers from the United States and Canada give an average of about .15,044 part in a thousand of mineral matter in solution, more than one-third being $CaCO_3$. The average amount of mineral matter in solution in 48 European streams cited by Bischoff[1] is .2127 part in a thousand, of which $CaCO_3$ is rather more than half. The average mineral matter in solution in 36 rivers cited by Roth[2] (including some of those tabulated by Bischoff) is .2033 part in a thousand, of which $CaCO_3$ is slightly less than one-half.

An average for American and European rivers, so far as determinable from data at hand, is about .1888 part in a thousand in solution, of which $CaCO_3$ is slightly less than one-half. These last figures are probably not very far from an average for river water in general.

The following table shows the total amount of solids carried in solution by the rivers indicated:[3]

Rhine	5,816,805	tons	per	year.
Rhone	8,290,464	“	“	“
Danube	22,521,434	“	“	“
Thames	613,930	“	“	“
Nile	16,950,000	“	“	“
Croton	66,795	“	“	“
Hudson	438,000	“	“	“
Mississippi	112,832,171	“	“	“

ECONOMIC CONSIDERATIONS.

Certain considerations of human interest in connection with river eorsion are worthy of note. When a drainage system has reached its mature stage its basin has the roughest topography which it will have at any time during that cycle of erosion. At that stage, therefore, road construction is relatively difficult. If the relief be great, roads must follow the valleys, or the crests of the ridges between them, if they would avoid heavy grades. In such regions roads are usually few and crooked.

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[1] Chemical Geology, Vol. I, pp. 76, 77, English ed., 1854.

[2] Allgemein und chemische Geologie, Vol. I, pp. 456, 457. 1879.

[3] Russell. Rivers of North America, p. 79.

The stage of development of valleys has an influence on the navigability of their streams. Streams well advanced in life are much more readily navigable than young ones, because their grades are lower and their volumes of water greater. Old streams, on the other hand, are sometimes depositing sand or silt along their lower courses to such an extent as to interfere with navigation.

At certain stages of their development the power of streams is more easily utilized than at others. Young streams, depending as they do for their supply on the rainfall of a limited area, are likely to be fitful in their flow, and therefore unreliable as a source of power. This is especially true where the precipitation is unequally distributed, and where the slopes are steep and free from forests. Because of their great volume, old and large streams, though sluggish, have great power, but it is less easily controlled. Where streams are large enough to be navigable industrial considerations often prevent the utilization of their power, the streams being more serviceable as highways than as sources of power. Other things being equal, it follows that streams are most available for water-power when they are large enough to have a moderately steady flow, and not so large as to be beyond ready control, or to be valuable for purposes of navigation.

Streams are subject to more disastrous floods in some stages of their development than in others. Floods resulting from heavy rains are likely to be greatest where the slopes above the drainage lines are on the whole greatest, for this is the condition under which the water is most quickly gathered into the drainage channels. The most disastrous floods, humanly speaking, are those which affect wide-bottomed valleys, where the flats are settled. In such cases a relatively slight rise may flood very extensive areas. In such valleys the most disastrous floods are generally in the spring, when the waters from the melting snows of the preceding winter are being discharged.[1] Many other considerations enter into the problem of floods. The presence of forests and other forms of vegetation on the slopes retards the flow of water into the valleys, and so tends to prevent floods, or at any rate to make them less severe. Porous soil and subsoil, or in their absence porous rock, absorb the rainfall, and prevent its prompt descent into the valleys and so tends to prevent or diminish floods.

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[1] For disastrous floods of the lower Mississippi, see Johnson, Bull. Geol. Soc. Am., Vol. II, pp. 20–25. For effect of precipitation and forests on floods, see Russell's Meteorology, pp. 198–217, and Vermeule, Report on Water Supply, Geol. Surv. of N. J.

The acreage of arable land within a given area stands in some relation to its drainage development. At an early stage in its erosion history, before an upland has been dissected by valleys, nearly all of it may be arable. Later, when drainage is at its maturity, and when hillsides and ridge slopes constitute a large part of the area, there is probably the least acreage of arable land. This is especially true if the slopes are so steep as to allow the soil to be readily washed away. At a still later stage, when the valley bottoms have become wide and the slopes of the ridges and hills so reduced as to be available, the area of cultivable land is again increased.

Marshes, ponds, and lakes have some bearing on the resources and industries of a region, and they stand in a more or less definite relation to the stage of erosion in which a region finds itself. In its youth ponds and lakes may occupy much of the surface; in its maturity they will have been largely drained.

These suggestions are sufficient to show that the topography of a region, even in so far as shaped by erosion, touches human interests at many points.

ANALYSIS OF EROSION.[1]

Erosion is the term applied to all the processes by which earthy matter or rock is loosened and removed from one place to another. It consists of three sub-processes, namely, *weathering*, *transportation*, and *corrasion*.

Weathering.

The term weathering is applied to nearly all those natural processes which tend to loosen or change the exposed surfaces of rock. The lettering of inscriptions on exposed marble becomes fainter and fainter as time goes by, and finally disappears, because the rock in which the letters were cut has weathered away. Some of it has crumbled off as the result of the expansion and contraction induced by changes of temperature, and some of it has been dissolved by the rain which has fallen upon it. In this case the weathering is effected partly by the atmosphere and partly by water. These are the chief, but not the only agents concerned in the general processes of weathering. Those phases of weathering which are the result of the activities of the atmosphere, whether physical or chemical, have been discussed in connection with the atmosphere (pp. 42 and 54).

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[1] An excellent discussion of this subject is given by Gilbert in The Henry Mountains, pp. 99 et seq., and more briefly in the Am. Jour. Sci., Vol. XII, p. 85 et seq. 1876.

The rain which falls upon the surface of exposed rock, and that which sinks through the soil to the solid rock below, dissolves, even if slowly, some of the rock constituents. Each constituent of a rock composed of several minerals may be looked upon as a binding material for the others. When one is dissolved the rock crumbles, much as mortar does when the lime which cements the sand is dissolved.

The solution of mineral matter by ground water, as well as the other chemical changes it effects, is greatly augmented by the impurities, especially carbonic and other organic gases, dissolved by the water from the atmosphere and the soil. The commonest chemical changes effected by the joint action of water and air, oxidation and carbonation, have been referred to in Chapter II. Hydration is more exclusively the work of water, and is one of the commonest processes of rock change, and often of rock disintegration. Numerous other less simple chemical changes resulting from the activities of ground water are constantly in progress, and in so far as they lead to the disintegration of rock are processes of weathering. Many chemical changes involve notable changes in volume of the mineral matter concerned. Merrill has calculated that in the conversion of the granitic rock of the vicinity of Washington, D. C., into soil, its volume has been increased 88 percent., largely as the result of hydration.[1] Even when the chemical changes do not themselves directly involve the disintegration of the rock, the accompanying increase of volume is sometimes sufficient to cause its physical disruption. This also may be regarded as a phase of weathering.

The weathering accomplished by water, or under its influence, proceeds at rates which vary with the composition of the rock, the amount and composition of the water, the temperature, and certain other factors less susceptible of brief statement. The weathering effected by ground water has a wider range both in area and depth than that due to changes of temperature, for while the latter is effective only where temperature changes are considerable, and where coherent material lies at the surface (p. 45), the former is operative to all depths to which water sinks.

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[1] Jour. of Geol., Vol. IV, p. 718. An excellent summary of the principles of Rock Weathering.

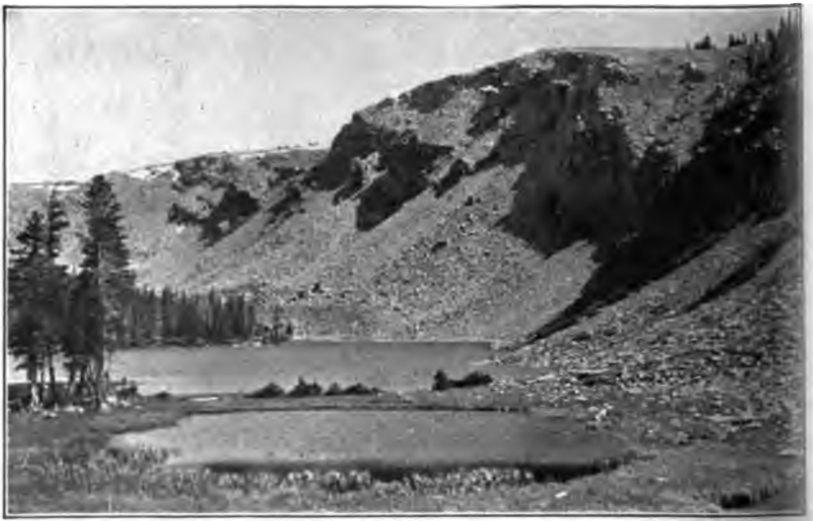


Fig. 94.—Talus accumulation at the base of a steep bluff. Weber Canyon, Uinta Mountains, Utah. The talus has accumulated since the last glaciation of the valley and is therefore of very recent origin. (Church.)

There are other processes of weathering not due directly either to the atmosphere or to water. The roots of trees and smaller plants frequently grow into cracks of rocks, and, increasing in size, act much like freezing water (p. 45) in similar situations. This wedge-work of roots is a phase of weathering.

From the faces of steep cliffs masses of rock frequently fall. However dislodged, their descent is effected by gravity. The quantities of débris at the bases of many cliffs, forming slopes of *talus* (Fig. 94) testify to the importance of the action of gravity in getting material from higher to lower levels. Another phase of gravity-work is shown in Fig. 95. Here, under the influence of gravity and expansion and contraction, due to freezing and thawing and wetting and drying the surface material is creeping down slope. In the process the rock is being broken. The process illustrated by the figure involves weathering as well as other factors.

The foregoing are among the commoner processes of weathering, although they do not exhaust the list. The more active and tangible processes by which surface rocks are broken up, such as wave wear, river wear and glacier wear, are processes of corrasion. The mechanical wear effected by wind-driven sand might be considered either as corrasion or as weathering. It is more likely to be regarded as corrasion if the amount of wear is considerable enough to be obvious. Rock is sometimes decomposed by the chemical action of hot vapors, gases, and waters rising to the surface from considerable depths. This is often seen in volcanic regions. A conspicuous illustration is seen in the canyon of the Yellowstone in the National Park. Decay of this sort is perhaps not properly weathering, but is not always readily distinguished from it.

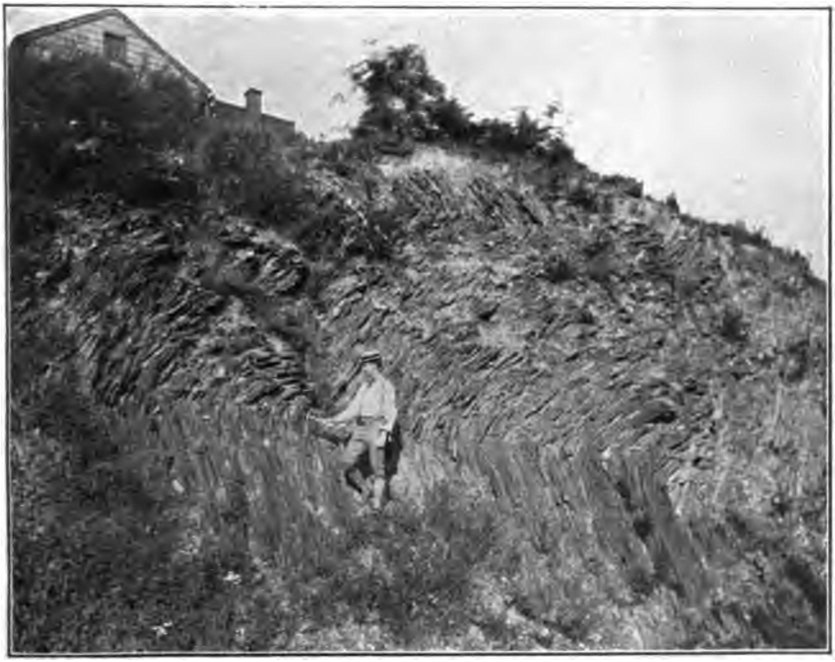


Fig. 95.—Shows the downward creep of soil and slaty rock under the influence of gravity.

The importance of weathering in the general processes of erosion is shown in many ways. In regions where the mantle rock is the product of the decay of the solid rock beneath, and such regions constitute a large portion of the earth's surface, the soil and subsoil represent the excess of weathering over transportation. Since most of the earth's surface is covered with soil to a greater or less depth, it is clear that, on the whole, weathering keeps ahead of transportation. Again, it is clear that the loosening of rock by weathering greatly increases the erosion which a given amount of moving water can accomplish. Not only this, but weathering plays a much more important rôle in the development of valleys than is commonly realized. This is best illustrated by the valleys of young swift streams. The valley which is not at its top ten times as wide as its stream is rare. The stream which has such a canyon has been cutting chiefly at its bottom. Ignoring its lateral corrasion, which is slight, the valley which it would cut would have a width equal to its own. This is illustrated by Fig. 96.

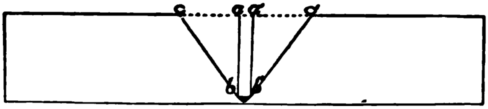


FIG. 96.—Diagram of a valley the top of which is ten times the width of the stream.

Weathering in its broadest sense is largely responsible for the width of such a valley, in so far as it exceeds the width of the stream. The work of weathering, slope wash, etc., has been to get the material which originally lay between *a*, *b*, and *c* down to the stream. The stream has then carried it away. The above illustration would not apply to old and sluggish streams, for they, by their meandering, widen their valleys independently of weathering.

Weathering is a part of erosion, but only a part. In so far as it is effected by solution the process involves the transportation of that which is dissolved to some other point. Transportation is also involved to some extent in the other processes of weathering, but the central idea of the processes embraced under this term is the loosening and disrupting of rock by which it is prepared for transportation.

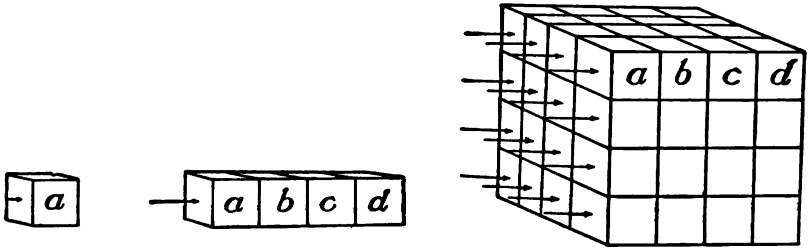
Transportation.

The second element of erosion is transportation. The transportation of mechanical sediment is to be distinguished from the transportation of materials in solution. In so far as mineral matter is dissolved it becomes, so far as flowage is concerned, a part of the stream. If the quantity dissolved were large it might influence the mobility of the water, but the amount is usually too slight to influence the flow sensibly.

The sediment transported by a stream is either rolled along its bottom or carried in suspension at some higher level. The coarser materials (gravel and sand) are carried chiefly in the former position, and the finer (silt and mud) largely in the latter.

Transporting power and velocity. — The transporting power of running water depends on its velocity. The formula expressing the relations between them is as follows: Transporting power, t, varies as the sixth power of velocity, v, ($t \propto v^6$); that is, doubling the velocity of the stream increases its transporting power 64-fold. Strictly speaking, this means that if a stream of given velocity is just able to move a stone of a given size, a stream with double that velocity will be just able to move a stone of the same shape 64 times as large as the first. This may be graphically illustrated as follows: Let a current be supposed just able to move the cube *a* (Fig. 97). If the current be doubled, twice as much water will strike the same surface with twice the force in the same time; that is, the force exerted on the cube *a* will be quadrupled. It will, therefore, be able not only to move the one cube, but it will be able to move three other cubes (*b*, *c*, and *d*) besides (Fig. 98). The same current against any other equal surface would also be able to move four small cubes, and there are sixteen such surfaces on the face of the large cube (Fig. 99). It follows that the dimension of the cube which the stream with the doubled velocity can move is four times as great as that of the cube which the original current could move, and the cubical contents of such a cube is 64 times as great as that of the first ($64 = 2^6$) (Fig. 99). Swift streams, therefore, have enormously greater power of transportation than sluggish ones. It does not necessarily follow that transportation keeps pace with transporting power; that depends on the accessibility of materials suitable for transportation. A stream of great transporting power, like the Niagara at its rapids, may carry little sediment, because there is little to be had.

The velocity of a stream depends chiefly on three elements—its gradient, its volume, and its load, (i.e., the sediment it is moving). The higher the gradient the greater the volume, and the less the load the greater the velocity. The relation between gradient and velocity is evident; that between volume and velocity is illustrated by every stream in time of flood, when its rate of flow is greatly increased. The relation between velocity and load is less obvious, but none the less definite. Every particle of sediment carried by a stream makes a draught on its energy, and energy expended in this way reduces the velocity. The draught on a stream's energy of a particle carried in suspension is measured by its mass into the distance it would fall in a unit of time in still water. It follows that a large particle makes a stronger draught on a stream's energy than the same amount of material in smaller pieces. It follows also that the comminution of sediment facilitates transportation in much more than a simple ratio, for not only can a given amount of energy carry more fine material than coarse, but a larger proportion of a stream's energy can be utilized in the transportation of the fine.



FIGS. 97–99.—Diagrammatic representation of the effect of increased velocity on transporting power.

How sediment is carried.—Coarse materials, such as gravel stones, are rolled along the bottoms of the swift streams which carry them. Their movement is effected by the impact of water. The same is true to a large extent of sand grains, especially if they be coarse. So far as concerns the material rolled along the bottom it is to be noted that a stream's transporting power is dependent on the velocity of the water at its bottom. This is much less than the surface, or even the average velocity. The particles of fine sediments, such as silt and mud, are frequently carried by streams quite above their bottoms, as shown by the roiliness of many streams. A particle of mud is usually a small bit of mineral matter, the specific gravity of which is two or three times that of water. Why does it not sink through the water and come to rest at the bottom of the stream, or suffer transportation as the gravel does?

A particle of sediment in running water is obviously subject to two forces, that of the current which tends to move it nearly horizontally down-stream, and that of gravity which tends to carry it to the bed of the stream. In Fig. 100, the arrows *ab* and *ac* represent respectively the relative force of gravity and a current of 5 miles per hour. As a result of these two forces the particle would tend to descend in the general direction of *ad*, a line which represents the resultant of these forces, though not the exact path which a particle acted on by them would take in water. If a river were the simple straightforward current which it is popularly thought to be, a particle in suspension would reach its bottom in the time it would take to sink through an equal depth of still water, for the descent would be none the less certain and none the less prompt because of the forward movement of the water. The current would simply be a factor in determining the position of the particle when it reached the bottom, not he time of reaching it. Very fine particles, like those of clay, though aving the same specific gravity as grains of sand, would sink less eadily than coarser ones, because they expose larger surfaces, relative o their mass, to the water through which they sink. But even such articles, unless of extraordinary fineness, would presently reach the ottom if acted on only by a horizontal current and gravity. Since ven sediment which is not of exceeding fineness is kept in suspension is clear that some other factor is involved. This is found, in part t least, in the subordinate upward currents in a stream.

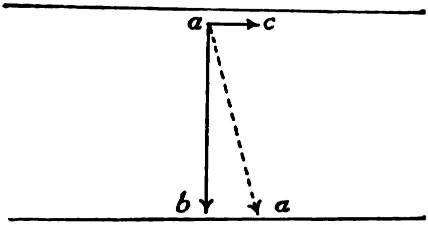


FIG. 100.—Diagram to illustrate the relative strength of the two forces acting on a particle in suspension. The arrows represent the relative strength of the two forces when the stream's velocity is 5 miles per hour. No account is taken in the diagram of the viscosity of the water, or of the acceleration of velocity of fall.

Where a bowlder occurs in the bed of a stream (Fig. 101) the water hich strikes it is in part forced up over it. If there be many bowlders e process is frequently repeated, and the number of upward currents is great. Any roughness will serve the same purpose, and every stream's bed is rough to a greater or less extent. Where there are roughnesses at the sides of a channel, currents are started which flow from them toward the center. The varying velocities of the different parts of a stream serve a similar purpose. The curves in a river tend to give the water a rotatory movement. A river is therefore to be looked upon not as a single straightforward current, but as a multitude of currents, some rising from the bottom toward the top, some descending from top to bottom, some diverging from the center toward the sides, and some converging from the sides toward the center. The existence of these subordinate currents is often evident from the boiling and eddying readily seen in many streams. It is, of course, true that the sum of the upward currents is always less than the sum of the downward, so that the aggregate motion of the water is down slope; but it is also true that minor upward currents are common. Sediment in suspension is held up chiefly by such currents, which, locally and temporarily, overcome the effect of gravity. The particles in suspension are constantly tending to fall, and frequently falling; but before they reach the bottom many of them are seized and carried upward by the subordinate currents, only to sink and be carried up again. Even if they reach the bottom, as they frequently do, they may be picked up again. It is probable that every particle of sediment of such size that it would sink readily in still water is dropped and picked up many times in the course of any long river journey, and its periods of rest often exceed its periods of movement.

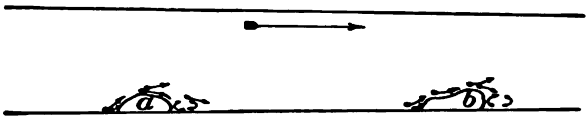


FIG. 101.—Diagram to illustrate the effect of bowlders, *a* and *b*, in a stream's bed on the currents of water impinging against them.

Independently of the subordinate currents, the different velocities of the different parts of a stream tend to keep materials in suspension by exerting different pressures on the different sides of suspended particles.[1]

River ice sometimes facilitates the transportation of débris which the water alone could not carry. The ice freezes to bowlders in the banks of the streams, to those which are partially submerged, and sometimes to those altogether submerged beneath slight depths of water. When the ice breaks up in the spring such bowlders, buoyed up by the ice, may be floated far down the stream. The influence of ice in this connection is most considerable in high latitudes, but it is of consequence as far south as Virginia, where the river deposits sometimes contain bowlders which the unaided streams could not have carried. *Ground ice* sometimes forms about bowlders in the bottoms of streams, especially in the quiet pools of turbulent rivers, and floats them to the surface before the surface itself is frozen.[1] In the floods of spring rivers often spread their ice widely over their flood-plains. It is sometimes massed in constricted portions of valleys so as to form great dams, the breaking of which is attended with great destruction.

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[1] Russell. Rivers of North America, p. 17.

Corrasion.

Abrasion.—The wear effected by running water is *corrasion.* So long as the materials to be carried away are incoherent it is easy to see how running water picks them up and carries them forward. The water which gathers in the depressions on the slope of a cultivated field gathers earthy matter from the surface over which it passes, even before it is concentrated into rills, and the rills continue the process. Thus the loose materials of the surface are gathered at the very sources of the streams, and the amount of sediment in the water after a heavy shower, even at the head of the stream, may be great. The run-off from the slopes of any valley in any part of its course likewise brings sediment to the stream, which gathers more from its bed wherever it flows with sufficient velocity over incoherent material. Streams also undercut their banks, and receive new load from the fall of the overhanging material.

By far the larger part of the sediment acquired by a normal stream is made up of material loosened in advance by the processes of weathering. The stream, or the waters which get together to make the stream, find them ready-made; but rivers frequently wear rock which is not weathered, for the principal valleys of the earth's surface are cut in solid rock, and many of them in rock of exceeding hardness. How does the stream wear the solid rock?

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[1] **W. G. Thompson. Nature, Vol. I, p. 555, 1870. The Matapediac River, N. B. Cited by Russell in Rivers of North America, p. 25.**

When a stream flows over a rock bed, the wear which it accomplishes depends chiefly on the character of the rock, the velocity of the stream, and the load it carries. If the rock be stratified and in thin layers, and if these thin layers be broken by numerous joints at high angles to the stratification planes, the impact of the water of a clear stream of even moderate strength may be effective in dislodging bits of the rock. This condition of things is often seen where streams run on beds of shale or slate. If the rock be hard and without bedding-planes and joints, or if its layers be thick and its joints few, clear water will be much less effective. If the surface of the rock be rough, the mechanical action of a swift stream of clear water might still produce some effect on it; but if massive hard rock presents a smooth surface to a clear stream, the mechanical effect of even a swift current is slight.

This general principle is illustrated by the Niagara River. Just above the falls the current is swift. When the river is essentially free from sediment, the surface of the limestone near the bank beneath it is sometimes distinctly green from the presence of the one-celled plants (fresh-water algæ) which grow upon it. The whole force of the mighty torrent is not able to sweep them from their moorings. Were the stream supplied with a tithe of the sand which it is capable of carrying, it would not take many hours, and perhaps not many minutes, to remove the last trace of vegetation. This illustration furnishes a clue to the method by which the erosion of solid rock in a stream's bed is effected.

It has been seen that the ingathering waters which make a stream often have abundant sediment before they reach well-defined stream channels, and that the streams continue to gather sediment wherever their beds are composed of material which is readily detached. The sediments which the stream carries are the tools with which it works. Without them it is relatively impotent, so far as the abrasion of solid rock is concerned; with them, it may wear any rock over which it passes (Fig. 102).

We have next to inquire the methods by which running water uses its tools in the excavation of valleys. When gravel is rolled along in the channel of a stream there is friction between it and the bed over which it moves. If the pebbles be as hard as the bed over which they are rolled their movement must result in its wear, and even if they be softer more or less wear takes place. As the moving stones wear the rock of the stream's bed they are themselves worn by impact with it and with one another. In all cases the softer material suffers the more rapid wear. The first effect of wear on materials in transportation is the reduction of their rugosities of surface. The projecting points and sharp angles are worn off, and the stones are reduced to rounded *water-worn* forms. The particles broken off make grains of sand, or, if very fine, particles of silt or mud. Even after a stone has been rounded it is subject to further wear and reduction, and in the course of time may be literally worn out.

The sediment carried in suspension, as well as that rolled along the bottom, may wear the rock bed of a stream. When a grain of sand in suspension escapes from an upward moving current it may not sink quietly. If it be caught by a downward current it may be made to strike a blow on the bed of the stream, and the effect of the blow is to wear the surface which receives it. The larger the grain and the stronger the current the greater the wear.

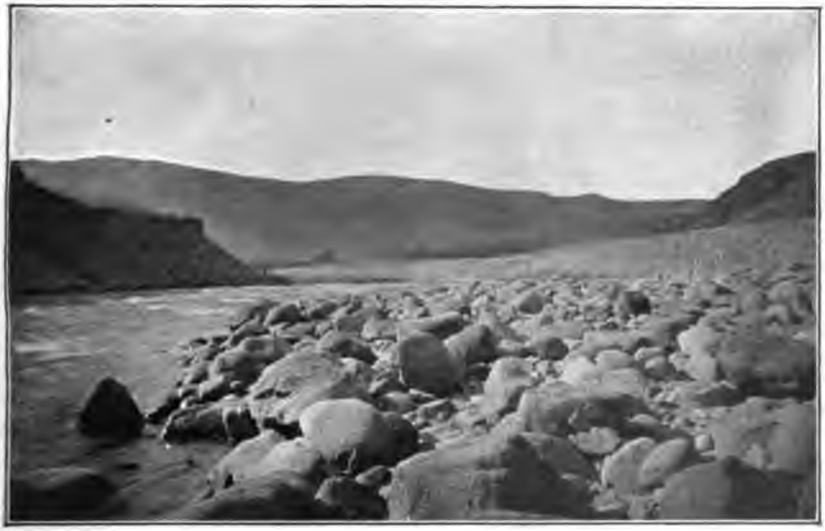


FIG. 102.—Some of the tools with which a stream works. The cobbles and bowlders have been shifted by the stream in its flow. Other stones and bowlders now in transit cause the ripples in the stream. The Chelan River, Wash., just above its junction with the Columbia. (Willis, U. S. Geol. Surv.)

The ceaseless repetition of the blows struck by the material in suspension, or rolled on its bottom, hour after hour, day after day, and year after year, will accomplish sensible results. In the long course of the ages this process has excavated deep valleys. Concomitant processes are largely concerned in making valleys wide, but the depth of valleys cut in solid rock is chiefly the result of the impact and friction of the sediment in transportation.

The wear effected in this way is not proportional to the number of blows struck. Since every pebble and every grain of sand carried diminishes the velocity of a stream, and since with diminished velocity the force of the blows struck is diminished, it follows that the blows may become so weak, as the result of their multiplication, as to be ineffective. The larger the load, therefore, which the stream carries, the more the tools with which it has to work, but the less effectively can it use them; and the load may be so far increased as to destroy its corrasive power altogether. On the other hand, the smaller the load of the stream the greater its velocity and the more effectively will its tools be used; but their number may be so far reduced that their aggregate effect is slight. To accomplish the greatest results on a bed of solid rock a stream must have tools to work with, but must not be so heavily burdened as to interfere with its effective use of them.

Whatever the cause of their unequal velocities swift and slow streams corrade their valleys differently. The erosion of a swift stream is chiefly at the bottom of its channel. The sluggish stream lowers its channel less rapidly, while lateral erosion is relatively more important. The result is that slow streams increase the width of their valleys more than the depth, while swift streams increase the depth more than the width. It follows that slow streams develop flats, while swift ones do not. Not only is a slow stream more likely to have a flat, and therefore a better chance to meander, but it is more likely to take advantage of opportunities in this line, for a slow stream gets out of the way for such obstacles as it may encounter, while a swift stream is much more likely to get obstacles out of its way.

Special phases of corrasion are introduced where waterfalls and other peculiarities dependent on inequalities of rock resistance occur.

Solution.—In most cases the solution effected by a stream is much less important than its mechanical work. Only when conditions are unfavorable to the latter, is solution the chief factor in the excavation of a valley. This may be the case where a stream's bed is over soluble rock, such as limestone, and where the stream is clear, or its gradient so low that its current is sluggish. The solvent power of water is not influenced by the presence of sediment, though the presence of sediment offers the water a greater surface on which to work.

CONDITIONS AFFECTING THE RATE OF EROSION.

In considering the rate of erosion, both the work of the stream in its valley and that of the general run-off are to be considered. The conditions which favor the most rapid erosion in a stream's channel are not necessarily those which determine most rapid degradation in the basin outside of the valley.

The Influence of Declivity.

In general the greater the declivity the more rapid the rate of erosion, whether in the stream's channel or on the slopes above it. The truth of this conclusion is illustrated by the great erosive power of swift streams as compared with slow ones.

It does not follow, however, that high declivity favors each element of erosion. The effect of declivity on weathering is far from simple. For example, great declivity, by allowing more of the rainfall to flow off over the surface, and by causing it to flow off more promptly, restricts the work of solution, and therefore of decomposition, both at the surface and beneath it. High declivity is also unfavorable to the growth of vegetation, and so to the wedge-work of roots. On the other hand, a given amount of wedge-work of roots and ice is more effective where the slope is steep than where it is gentle, for such materials as are loosened descend the slopes more readily. The prompt removal of weathered materials, by exposing fresh surfaces of rock, accelerates weathering. The total amount of weathering may therefore not be diminished by the increase of slope, even though certain of its processes are hindered.

The effect of high declivity on transportation, the second element of erosion, is too patent to need explanation.

Corrasion likewise is favored by high declivity, for the abrasive power of a stream increases as the square of its velocity. With corrasive power increased, corrasion will also be increased if the water has tools to work with. Since high declivity greatly increases both the transporting and the corrasive power of running water, and favors certain elements of weathering, it is clear that the aggregate effect of high declivity is to favor erosion, whether in the channel of the stream or on the general surface of its drainage basin.

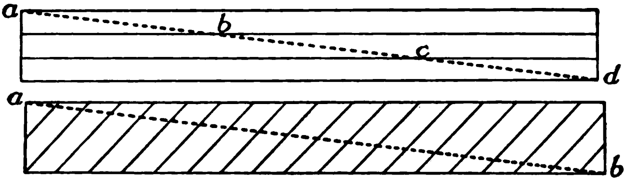
The Influence of Rock.

The physical constitution, the chemical composition, and the stratigraphy of a rock formation, influence the rate at which it may be broken up and carried away. Clastic or fragmental rocks are usually stratified and made up of cemented pebbles (conglomerate), sand grains (sandstone), or particles of mud (shale). Igneous rocks, such as granite, are massive instead of stratified, and are usually made up of great numbers of interlocking crystals which bind one another together. Some crystalline rocks, such as schists, though not stratified, possess cleavage, which has much the effect of stratification, so far as erosion is concerned. All rocks are affected by systems of more or less nearly vertical cracks called *joints*. All these structures have their influence upon the rate of degradation.

Physical constitution.—Clastic rocks may be firmly cemented, or their constituents may be loosely bound together. The less the coherence the more ready the disintegration, and the finer the particles the more easily are they carried away. When the particles in transportation are angular they effect more wear on the bed over which they move, and on one another, than when they are round. The difference is great where the particles are large, and little where they are very small. If the materials carried be harder than the bed over which they pass, corrasion of the latter is favored.

Chemical composition.—Something also depends on the chemical composition of the rock, since this affects its solubility, and therefore its rate of decomposition. The more soluble the rock the larger the proportion of it which will be taken away in solution; but it does not follow that the most soluble rock will be most rapidly eroded, since the rate of erosion depends on abrasion as well as solution, and a rock which is readily soluble, as rocks go, may be less easily abraded than a rock which is made of discrete and insoluble particles bound together by a soluble cement. In such rocks, for example a sandstone in which the grains are cemented together by lime carbonate, the solution of the cement sets free a considerable quantity of sand, so that a small amount of solution prepares a large amount of sediment for removal. A stream might cut its valley much more rapidly in such a sandstone than in a compact limestone, though the latter is, as a whole, the more soluble. The constituent minerals of crystalline rocks resist solution and decay unequally, and when any one is dissolved or decomposed the rock crumbles and the less soluble constituents are ready for removal by mechanical means. So long as the material loosened by disintegration is removed, chemical heterogeneity favors erosion; but if the loosened débris is not removed erosion is not favored by chemical heterogeneity. In such a case erosion would be most rapid where the rock was most soluble.

Structure.—The structure of the rock has much to do with the rate of its erosion. Other things being equal, stratified rock is more readily eroded than massive rock, since stratification-planes are planes of cleavage, and therefore of weakness. Taking advantage of these planes the water has less breaking to perform to reduce the material to a transportable condition. For the same reason a thin-bedded formation is more easily eroded than a thick-bedded one.



Figs. 103 and 104.—Diagrams to illustrate the fact that a stream crosses many more cleavage-planes when the beds of rock are inclined than when they are horizontal.

The beds of stratified rock may be horizontal, vertical, or inclined, and inclined strata may stand at any angle between horizontality and verticality. In indurated formations the rate of erosion is influenced both by the position of the strata and by the relation of the direction of the flowing water to their dip and strike. On the whole the strata which are horizontal, or but slightly inclined, are probably less favorable for rapid erosion than those which are vertical or inclined at considerable angles. This is at least true where the layers are of uniform hardness and the joints infrequent.

Horizontal strata expose fewer cleavage planes to the water flowing over them than strata in any other position. In Fig. 103 the stream which has the profile *ad* crosses bedding-planes at *b* and *c*. In Fig. 104, where the beds dip up-stream, many more division-planes are crossed in the same distance. Since bedding-planes are planes of weakness, it follows that horizontal and nearly horizontal strata are not, under ordinary conditions of erosion, in a position favorable for most rapid wear. When strata are horizontal, it makes no difference which way the stream runs, for the current sustains the same relation to the cleavage-planes whatever its course.

In the case of incoherent material the position of the beds, or even their existence, has little influence on the rate of erosion. Such formations are weak in all directions, not simply along bedding-planes.

When the strata are vertical, three distinct cases may arise (Fig. 105). The stream may flow (1) with the strike (*aa*); (2) at right angles to the strike (*bb*); or (3) oblique to it (*cc*) at any angle whatsoever. It is perhaps not possible to say which of these positions is most favorable for erosion, for the character of the rock, the thickness of its layers, its ability to stand with steep slopes, and the strength of the currents concerned, would influence the result. A stream which flows at right angles to the strike (*bb*, Fig. 105) would cross more cleavage-planes in a given distance than a stream flowing in any other direction, and would strike the outcropping edges of layers at the angle of greatest advantage. A stream flowing along the strike (*aa*), on the other hand, has better opportunity to sink its channel on cleavage-planes, and the current oblique to the strike (*cc*), has some of the advantages of each of the others.

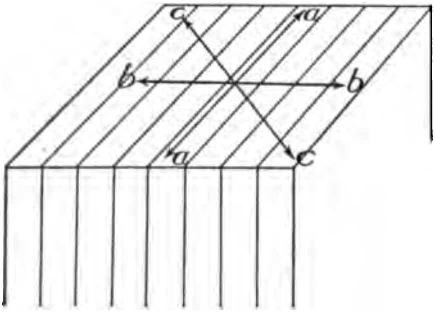


Fig. 105.—Diagram to illustrate the various relations a stream may sustain to the outcrop of vertical layers of rock.

When the strata are inclined five cases may arise. (1) The stream may be parallel to the strike (*aa*, Fig. 106), when it makes no difference which way the current flows; it may be at right angles to the strike (*bb'*), and (2) flowing with the dip (toward *b'*), or (3) against it (toward *b*); it may be oblique to the strike, and flowing (4) in the general direction of dip (toward *c'*); or (5) in the opposite direction (toward *c*). As before, the stream flowing at right angles to the strike would cross the largest number of layers in a given distance, and so have an opportunity to take advantage of more cleavage-planes than a stream in any other position. But in the case of inclined strata a new element enters into the problem. When the stream flows parallel to the strike, the valley which is in process of deepening is not sunk vertically, but is shifted more or less in the direction of the dip (Fig. 107). This is called *monoclinal shifting.* The result is that there is a constant tendency to undermine (sap) the valley bluff on the down-dip side, and this process of sapping will, according to its rate, accelerate the growth of the valley, especially in width. Monoclinal shifting is favored by the presence of a hard layer (*H*), as shown in Fig. 107, if this stratum is the bed of the stream.

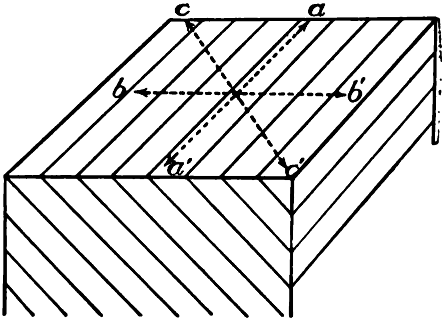


Fig. 106.—Diagram to illustrate the various relations a stream may sustain to the outcrops of inclined layers of rock.

In the second and third cases mentioned above, the only difference is in the angle at which the current strikes the outcropping edges of layers and laminæ. The mechanical advantage is with the stream which flows with the dip. In the fourth and fifth cases something will depend on the angle which the stream's course makes with the strike. In all these cases, as in those where the strata are vertical, much will depend on the thickness and resistance of the layers and on the strength of the currents concerned.

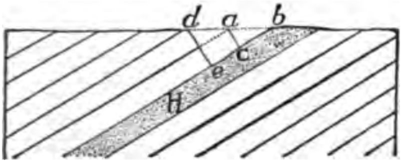


Fig. 107 —Diagram to illustrate monoclinal shifting. The valley *abc*, as seen in cross-section, becomes *deb*, as the stream lowers its channel.

The Influence of Climate.

Climate has both a direct and an indirect effect on erosion. Its direct influence is through precipitation, evaporation, changes of temperature, and wind; its indirect, through vegetation. Like declivity and rock structure, climate does not affect all elements of erosion equally.

The chief elements of climate are temperature, moisture, and atmospheric movements; the principal factors which influence it are latitude, altitude, distance from the sea, direction of prevailing winds, and topographic relations.

The effects of variations in temperature on rock weathering have already been discussed (p. 43). They are chiefly mechanical, and are seen at their best where the daily range is great.

High temperature favors chemical action, and the weathering of rock by decomposition is at its best in the presence of abundant moisture in regions where the temperature is uniformly high. Furthermore, a warm moist climate favors the growth of vegetation, the decay of which supplies the water with organic acids which greatly increase its solvent power. The climatic conditions favoring mechanical weathering are therefore different from those favoring chemical weathering. High temperature and abundant moisture and vegetation are found in many tropical regions, and here the rock is often decomposed to greater depths, on the whole, than in high latitudes. How far this is the result of rapid weathering, and how far of slow removal, due in part to the protective influence of the plants, cannot be affirmed. If the weathered material is not removed, it will presently become a mantle thick enough to retard the processes which brought it into existence.

So long as the water of the surface and that in the soil remains unfrozen, temperature affects neither corrasion nor transportation. But in middle and high latitudes the surface is frozen for some part of each year. During this time corrasion is at a minimum, for although the streams continue to flow there is relatively little water running over the surface outside the drainage channels, and that little is relatively ineffective. Under some conditions, therefore, temperature affects both corrasion and transportation.

The humidity of the atmosphere has an influence even more important than that of temperature on the rate of erosion, and its influence is exerted on each of the elements of that complex process. A moist atmosphere favors oxidation, carbonation, hydration, and the growth of vegetation, all of which promote certain phases of rock weathering. On the other hand, humidity tends to prevent sudden and considerable variations in temperature, thus checking the weathering effected by this means. Precipitation, the most important single factor in determining the rate of erosion, is dependent on atmospheric humidity. Its amount, its kind (rain or snow), and its distribution in time, are the elements which determine its effectiveness in any given place.

Other things being equal the greater the amount of precipitation the more rapid the corrasion and transportation. Much, however, depends on its distribution in time. A given amount of rainfall may be distributed equally through the year, or it may fall during a wet season only. The maximum inequality of distribution would occur if all the rainfall of a given period were concentrated in a single shower. With such concentration the volume of water flowing off over the surface immediately after the down-pour would be greater than under any other conditions of precipitation, and since velocity is increased with volume, and erosive power with velocity, it follows that the erosive power of a given amount of water would be greater under these circumstances than under any other. Furthermore, a larger proportion of the precipitation would run off over the surface under these circumstances than under any other, for less of it would sink beneath the surface and less would be evaporated. If erosive power and rate of erosion were equal terms, this would therefore be the condition for greatest erosion; but erosive power and rate of erosion do not always correspond. If the water falling in this way could get hold of all the material it could carry, extreme concentration of precipitation would be the condition favorable for most rapid erosion. But if the amount of available material for transportation is slight, a large part of the force of the water could not be utilized in erosion. It follows that if there were a large amount of disintegrated material on the surface, erosion would be greater the greater the concentration of precipitation. If, on the other hand, there were but little disintegrated material on the surface, frequent showers, with intervening periods when conditions were favorable for weathering, that is, for preparing material for transportation, might be more favorable for rapid erosion. While the total energy of running water available for erosion under these conditions would be less than before, there might in the long run be more material for transport; for weathering in the presence of moisture, and all that goes with it, might be more effective in preparing material for transportation, than weathering during the long periods of drought which would occur if the precipitation were concentrated to its maximum. Temperature favoring, the uniform distribution of moisture through the year would allow the growth of vegetation, which, although favoring some processes of weathering, retards erosion in general. While therefore it is not possible to say what distribution of rainfall favors most rapid erosion without knowing the nature of the surface on which it is to fall, enough has been said to show that the problem is by no means a simple one. Some of the most striking phases of topography developed by erosion, such as those of the Bad Lands (Figs. 75 to 78, and 108), are developed where the rainfall is unequally distributed in time, and too slight or too infrequent to support abundant vegetation.

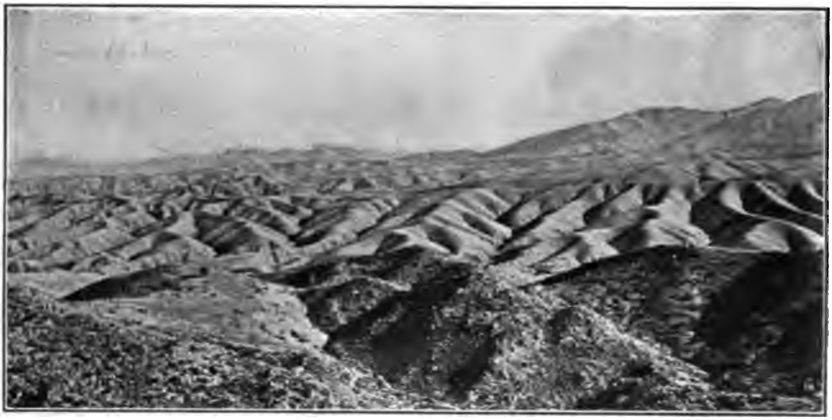


Fig. 108.—Bad-land topography developed under conditions of aridity and unequal distribution of rainfall. Slope of Pinal Mountains, Ariz. (Ransome, U. S. Geol. Surv.)

During its fall, and immediately after, rain is more effective than an equal amount of snow; but the snow may be accumulated through a considerable period of the year, and then melted rapidly, when it has an effect comparable to that which would be produced by the concentration of the rainfall into a limited period of the year. If the ground beneath be frozen when the snow melts (and this is often the case) the erosion accomplished by the resulting water will be diminished.

Except in dry regions, where wind-work sometimes exceeds water-work, the movements of the atmosphere are of less importance directly than precipitation in determining the rate of erosion. But even in regions which are not arid the winds have much to do with the rate of evaporation and the distribution of rainfall, so that their indirect effect is great. Even their direct effects in moist climates are not to be lost sight of, for even here the surface is sometimes dry enough to yield dust and sand, and the uprooting of trees so disturbs the surface as to make earthy débris more accessible to wind and water. Where trees gain precarious footholds on steep slopes, as they often do, they likely to be overturned as soon as they are large enough to offer considerable resistance to the wind, and in the overturning, large quantities of rock are sometimes loosened and carried down the slope by gravity. This phase of destructive work is seen at its best on the walls of gorges, where trees often flourish until their tops project above the rim of the valley.

Through vegetation, climate influences erosion in ways which are easily defined qualitatively, but not quantitatively. Both by its growth (wedge-work of roots) and by its decay (supplying CO_2, etc., to descending waters) it favors certain phases of weathering; but, on the other hand, it retards corrasion and transportation both by wind and water. This is well shown along the banks of streams and on the faces of cliffs, in clay, sand, etc. Its aggregate effect is probably unfavorable to erosion by mechanical means, and favorable to that by chemical processes.

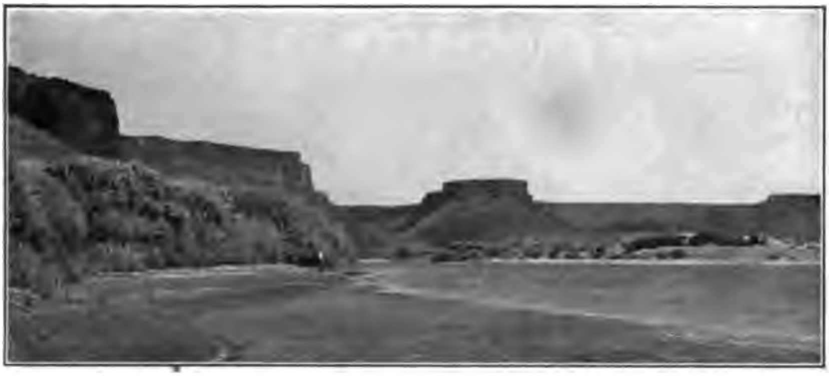


Fig. 109.—Characteristic cliffs of high arid regions. Right wall of Snake River canyon, nearly opposite the mouth of Salmon River, Id. Two spring-formed coves, with "Castle Rock" between. (Russell, U. S. Geol. Surv.)

Erosion in high arid regions differs from that in regions of abundant rainfall in several ways. It is obvious that the valleys will develop more slowly in the former, that they will remain young longer, that the period necessary for the dissection of the surface is greater, that the watercourses will be less numerous, and that fewer of them will have permanent streams. There are certain other differences which are less obvious. If the arid region be high and composed of heterogeneous strata, the topography which erosion develops is more angular (Fig. 83) than that of the humid region. This is because there is less rock decay, and less vegetation to hold the products of decay. The more resistant beds of rock therefore come into greater prominence, especially on slopes, where they develop cliffs (Figs. 109 and 110). These general principles find abundant illustration in the plateaus of the western part of the United States,[1] where the cliffs are by no means confined to the immediate valleys of the streams (Fig. 1, Pl. XII).

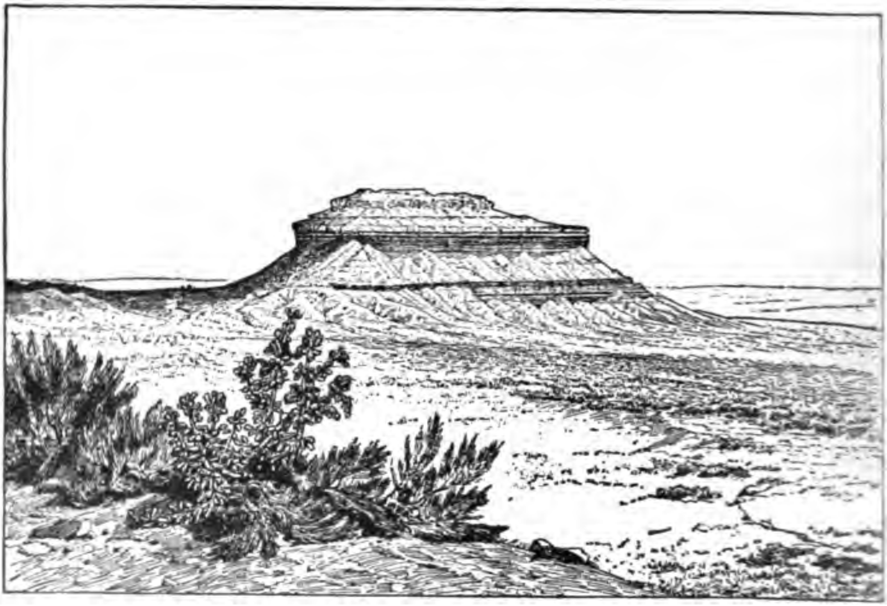


FIG. 110.—A Butte. A characteristic feature of the arid plateau region of the West. (Dutton, Mono. II, U. S. Geol. Surv.)

EFFECTS OF UNEQUAL HARDNESS.

In the preceding pages incidental reference has been made to the results of inequalities of rock resistance. This topic will now be considered more fully.

Rapids and falls.—Returning for a moment to the hypothetical island with which our study of erosion began, let a horizontal layer of hard rock be assumed to run through it (H, Fig. 111). As the rain falls on the land and runs off over it, wear will be less rapid where the hard layer comes to the surface than at the higher or lower levels. As a result, the slope will become steeper at and below the outcrop of the hard layer, and less steep immediately above it, as shown by *ab* in Fig. 111. Under these conditions the water passing over the hard ledge constitutes *rapids*. The increased erosion which accompanies the increased velocity makes the rapids more rapid. The process may continue until the water *falls*, rather than *flows* over the hard layer (*cd*, Fig. 111). With continued rainfall the edges of the hard layer, together with the slopes above and below, would continue to recede toward the center of the island. Under conditions of absolute homogeneity of material, save for the hard layer specified, no valley would be developed, and therefore no stream.

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[1] Dutton. Tertiary History of the Grand Canyon District, Mono. II, U. S. Geological Survey.

If the surface was so changed as to allow of the development of a valley (p. 63) the same principles would be applicable. As an active stream passes from a hard layer to one less resistant, the greater wear on the latter gives origin to rapids. At first the rapids would be slight (*a*, Fig. 112), but would become more considerable (*b*) as time and erosion go on. When the bed of the rapids becomes sufficiently steep, the rapids become *falls*[1] (*cd*). When the water falls rather than flows over the rock surface below the hard layer, erosion assumes a new phase. The hard layer is then undermined, and the undermining causes the falls to recede. This phase of erosion is sometimes called *sapping*.

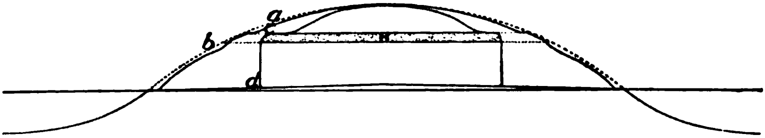

Fig. 111.—Diagram representing a horizontal layer of hard rock in an island, and its effects on erosion.

If the hard layer which occasions a fall dips up-stream (Fig. 112), its outcrop in the stream's bed becomes lower as the fall recedes (*c*). When it has become so low that the water passing over it no longer

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[1] The terms rapids, falls, and cataracts are rather loosely used. Many moderate rapids are incorrectly called falls. The "Falls of the Ohio" is an example. The term cataract is often applied to very steep rapids or falls.

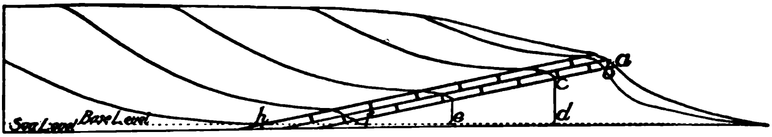

Fig. 112.—Diagram illustrating the development of a fall where the hard layer dips gently up-stream.

reacts effectively against the less resistant material beneath (*f*), sapping ceases, and the point of greatest erosion may be shifted from the soft material beneath the fall to the hard layer itself. The actual rate of erosion at this point may be no greater than before, though the relative rate is. Under these circumstances the vertical edge of the hard layer will presently be converted into an incline (*f*), and as this takes place the fall becomes rapids. The conversion of the falls into the rapids begins about the time the lower edge of the hard stratum in the channel reaches grade. By continuation of the process which transformed the falls into rapids, the rapids become less rapid, and when the upper edge of the hard layer has been brought to grade, the rapids disappear (*h*, Fig. 112). The history of rapids which succeed falls is the reverse of that which preceded. The later rapids are steepest at the beginning of their history, the earlier at their end. Stated in other terms, rapids are steepest when nearest falls in time. Slight differences in hardness in successive layers often occasion successive falls or rapids (Fig. 114).

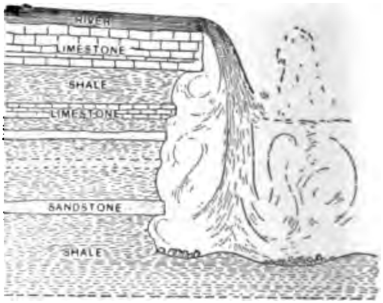

Fig. 113.—Diagram illustrating the conditions which exist at Niagara Falls. (Gilbert.)

If the hard layer which occasions the falls be horizontal, instead of dipping up-stream, the general result would be the same; but, other things being equal, the duration of the falls developed under these conditions would be greater, since they must recede farther before becoming rapids.

If the layers of unequal hardness in a stream's bed be vertical and the course of the stream at right angles to the strike, rapids, and perhaps falls, will develop (Fig. 115). The chances for falls are greater, the greater the difference in hardness. Falls developed under these conditions, as well as the rapids preceding and following, would remain constant in position until the resistant layer was brought to grade, but they would ultimately disappear as in the preceding cases. Falls are not likely to develop where the strata of the stream's bed dip down-stream, though they may develop even under these conditions if the gradient of the stream is greater than the dip of the strata (Fig. 116).

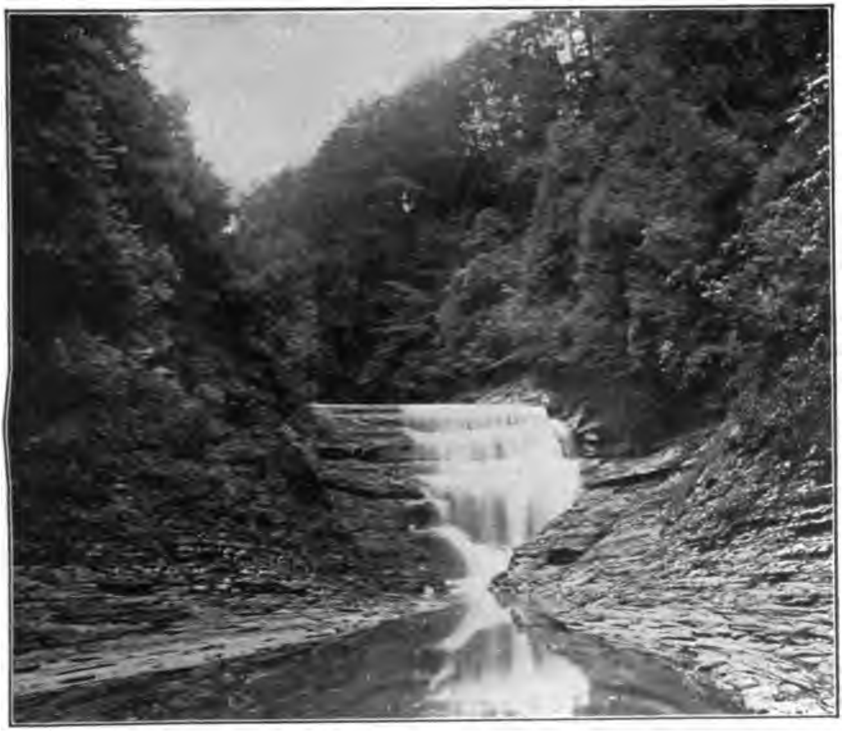

Fig. 114.—Falls in Utica shale, Canajoharie, N. Y. (Darton, U. S. Geol. Surv.)

The inequality of resistance in the rock which occasions a fall may be original or secondary. In the case of Niagara Falls[1] (Fig. 113) relatively resistant limestone overlies relatively weak shale. At the Falls of St. Anthony (Minneapolis) limestone overlies friable sandstone. The falls of the Yellowstone and the Shoshone Falls of the Snake River (Idaho), are in igneous rock. In the former case the unequal resistance is occasioned by unequal decay of the rock, due perhaps to the rise of hot vapors which have decomposed the rock along the lines of their ascent; in the latter, a more resistant sort of igneous rock overlies a less resistant.

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[1] **Gilbert, article on Niagara Falls, in Physiography of the United States.**

Structural features, such as jointing, sometimes give rise to falls, or determine their distinctive features (Fig. 117), even where the formations involved are of uniform hardness. A joint plane has the effect of a weak vertical or highly inclined bed. If an open joint is discovered in a stream's bed, the water enters it. If it finds an outlet below, a channel is worn along the new line of flow, with rapids or falls where the water descends. Rock originally homogeneous may be much fractured in some parts, while it remains unbroken in others. Where a stream passes from the solid to the broken portion rapids, or even falls, may develop.

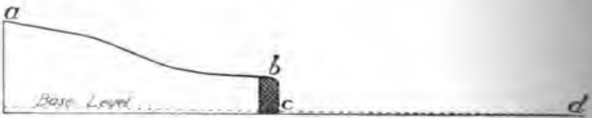

FIG. 115.—Diagram illustrating the development of falls over a vertical hard lay-

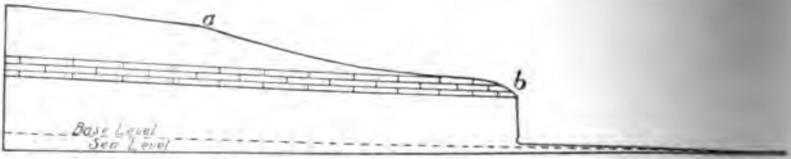

FIG. 116.—Diagram illustrating the possibility of falls where the beds dip down-stream.

Falls may originate in still other ways. If for any reason a stream is forced out of its valley, it may in its flow find entrance to another valley, or to another part of its own valley, over a steep slope. If the structure of the slope favors, a fall may speedily develop. The Falls of St. Anthony are an example, the Mississippi having been turned out of its earlier course by deposits of glacial drift. Again, if an obstruction of any sort, such as a flow of lava, dams a stream, rapids or falls are developed where the water overflows the dam. When a main valley is notably deepened by glaciation the drainage from tributary valleys may fall into it, if the tributaries were not equally deepened. Falls which originated in this way are common in the western mountains of the United States, as well as in most mountain regions recently affected by local glaciers (Fig. 118).

One waterfall often breeds others. Thus where a fall recedes beyond the mouth of a tributary stream, the tributary falls. The Falls of Minnehaha, on a small tributary to the Mississippi, near Minneapolis, may serve as an illustration. In such cases the falls may not develop from rapids. Once in existence, the fall of a tributary follows the same history as that of a main stream.

Streams which have falls are relatively clear.[1] If a stream favorably situated for the development of a fall carried a heavy load, deposition would take place below the rapids, and the tendency would

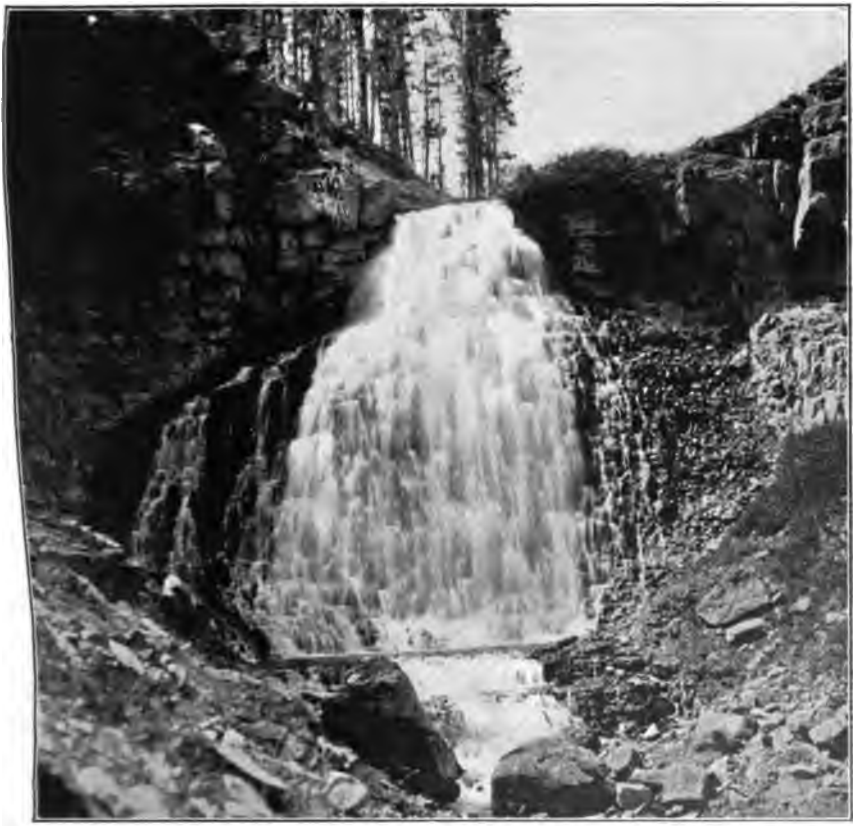

FIG. 117.—Kepler's Cascade, in the Yellowstone Park. The jointed and fractured character of the igneous rocks occasions a series of falls and rapids. (Iddings, U. S. Geol. Surv.)

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[1] Gilbert. Am. Jour. Sci., Vol. XII. p. 99, 1876.

be to aggrade the channel at that point and so to prevent the development of the fall. Falls occur only on streams which have relatively

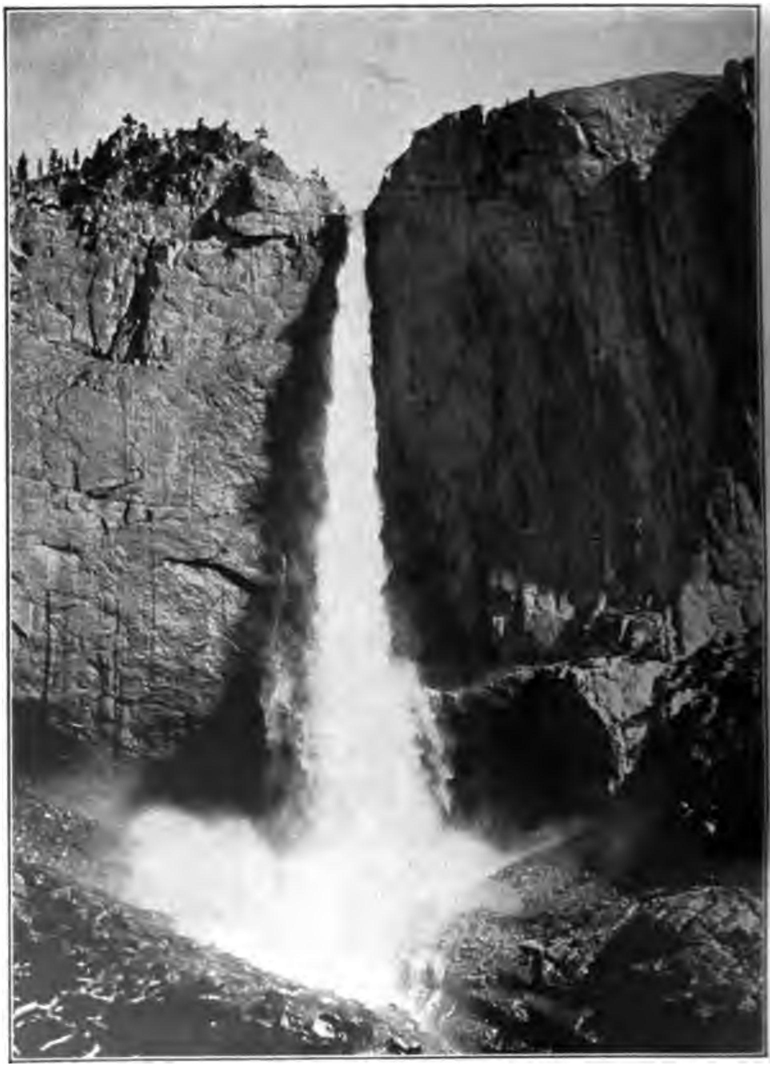

FIG. 118.—The Upper Yosemite Falls.

high gradients. This means that the streams which have falls ar well above base-level, and streams well above base-level are young. Falls therefore are a mark of topographic youth.

The fall of the Niagara[1] (Pl. IX) is one of the most remarkable known, both because of its large volume of water and its great descent,

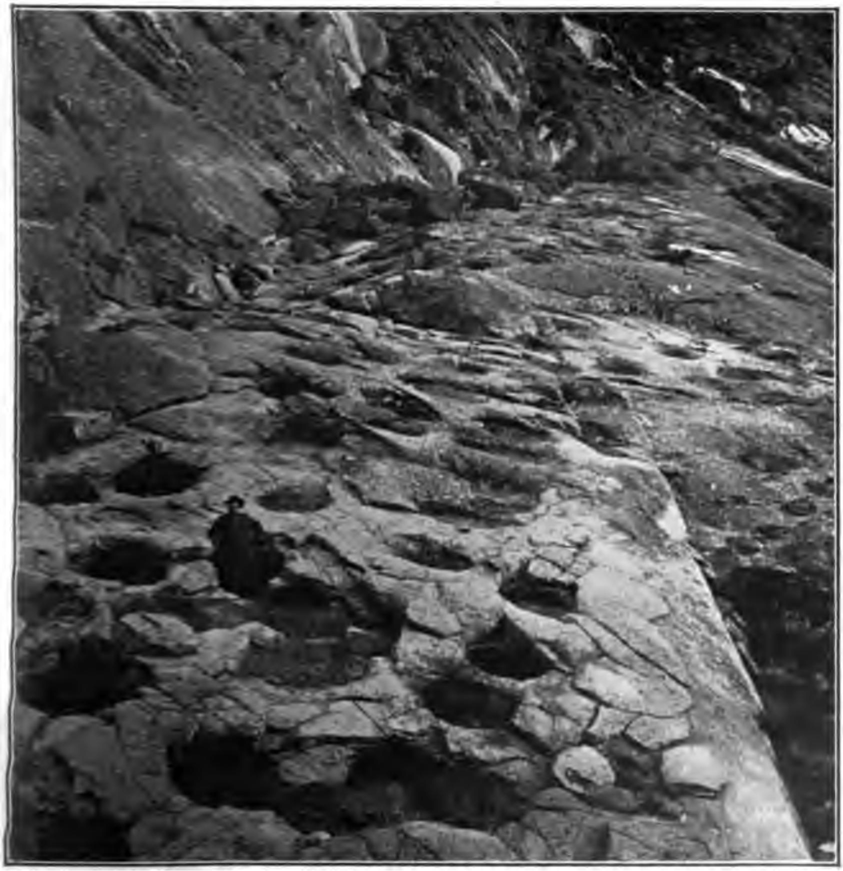

Fig. 119.—A group of pot-holes. (Turner, U. S. Geol. Surv.)

etween 160 and 170 feet. The rate at which the fall is receding is matter of interest not only in itself, but because, once determined, may be made to serve as a unit of measurement for certain important vents in geological history. It was formerly conjectured that this ll was receding at the rate of one to three feet per century, but it as not until recent years that its actual rate of recession was approximately fixed. By surveys executed in 1842 and 1890 it has been determined that its average rate of recession between those dates was something like 4½ feet per year, or about 150 times as great as the highest estimate stated above. It is to be noted that this is the *average* rate of recession, for all parts of the ledge over which the water falls are not receding at the same rate. The point of the "Horseshoe" has, during the same time, gone back at more than twice this rate.[1]

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[1] For a brief account of this fall see Gilbert in Physiography of the United States.

Rapids and falls sometimes occasion the development of *pot-holes* (Fig. 119), a peculiar rather than important erosion feature. The holes are excavated in part by the falling and eddying of silt-charged water, but chiefly by stones which the eddies move. Pot-holes which are not now in immediate association with rapids or falls often point to the former existence of rapids or falls.

Rock terraces.—The tendency to sapping shown in many waterfalls is also shown in the weathering and erosion of the sides of a valley where a hard layer outcrops above the bottom, and the profile of the side slopes of the valley simulates that of the stream; that is, the slope becomes gentle just above the hard layer, and steep, or even vertical, at and below its outcrop. This is illustrated by Fig. 120, where the hard layer through which the stream has sunk its valley stands out as a rock terrace on either side of the valley. Such terraces are not rare and are popularly believed to be old "water-lines"; that is, to represent the height at which the water once stood. In one sense this interpretation is correct, since a river has stood at all levels between that of the surface in which its valley started, and its present channel, but the shelf of hard rock does not mean that the river, after attaining its present channel, was ever so large as to fill the valley to the level of the terrace. Rock terraces may also result from changes of level.

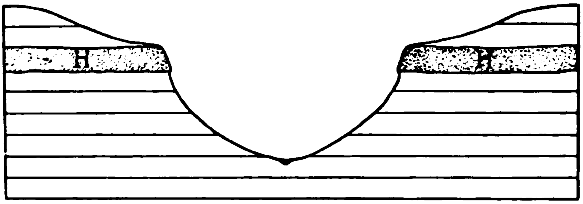


FIG. 120.—Diagram to illustrate the development of rock terraces.

[1] Gilbert. Science, Vol. VIII, p. 205, 1886.

Narrows.—Inequalities in hardness occasion another peculiarity common to valleys. If a stream crosses vertical or highly inclined strata of unequal hardness, its valley is usually constricted at the crossing of the harder layers. If such a constriction be notable it is called a *narrows*, or sometimes a *water-gap* (Figs. 121, 159, and Fig. 2, Pl. XII). The Appalachian Mountains afford numerous examples. The constriction arises because the processes which widen the valley are less effective on the hard layer than on the less resistant ones on either hand. Though most narrows are due to the superior resistance of the rock where they occur, they are sometimes the result of other causes

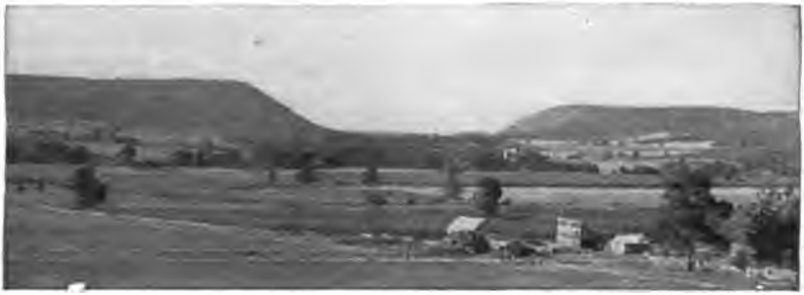


FIG. 121.—Lower narrows of the Baraboo River, Wis. The even-crested ridge is Huronian quartzite. The surroundings are of Cambrian sandstone. (Atwood.)

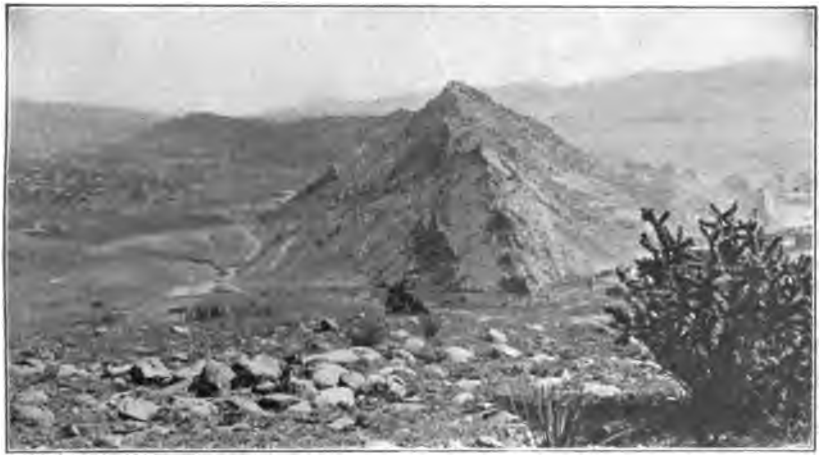


FIG. 122.—A hog-back, Jura-Trias. Colorado City, Colo. (Russell, U. S. Geol. Surv.)

Narrows are much more conspicuous in certain stages of erosion than in others. While a valley is still so young as to be narrow at all points, no narrows will be conspicuous; but at a later stage in its history, when the valley is otherwise wide, narrows are more pronounced. At a still later stage, when the hard strata themselves approach base-level, the narrows again become inconspicuous.

From what has preceded it is clear that rapids or falls are likely to occur at narrows, especially in the early part of their history.

Other effects on topography.—Inequalities in the hardness of rock develop certain peculiarities of topography other than those of valleys. The less resistant portions of a land area more or less distant from streams are worn down more readily than those which are more resistant. If great areas of high land be capped with hard rock they are likely to remain as plateaus after surrounding areas of less resistance are brought low. If the hard capping affects a small area instead of a large one, the elevation is a butte, a hill, or a mountain, instead of a plateau (Fig. 110). Many buttes and small mesas are but remnants of former plateaus (Mesa Lauriano, N. M., Fig. 1, Pl. XII). A feature of buttes and mesas capped by hard rock is the steep slope or cliff corresponding to the edge of the hard bed (Figs. 78 and 109).

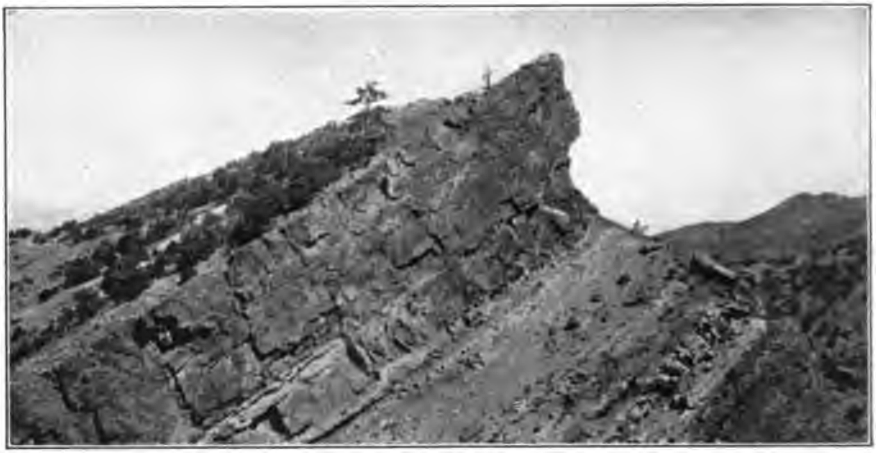


FIG. 123.—A ridge due to the outcropping edge of hard Jurassic rock. Wyoming.

If the rock of a region be stratified and the layers tilted, the removal of the softer beds leaves the harder ones projecting above the general level in the form of ridges or "hog-backs" (Figs. 122 and 123).

Dikes of igneous rock, harder than the beds which they intersect, likewise become ridges after the degradation of their surroundings. The plugs of old volcanic vents and other igneous intrusions of limited area often constitute conspicuous hills or mountains after erosion has removed their less resistant surroundings (Fig. 124). Inequalities of hardness are therefore responsible for many hills and ridges. In the isolation of the hills and ridges picturesque coves are developed, where the attitude and distribution of the weak and strong rocks are propitious. The bottoms of the coves are located on the weak rocks, and above them rise the precipitous slopes of the resistant ones. Round valley (Fig. 1, Pl. XVII, High Bridge, N. J., quadrangle, U. S. Geol. Surv.) and the coves about the head of Hiawassee River (Dahlonega, Ga., quadrangle) are examples.

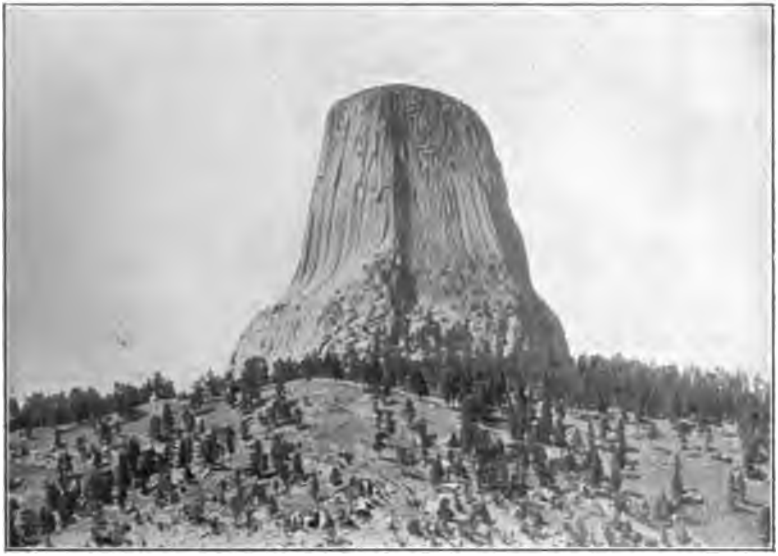
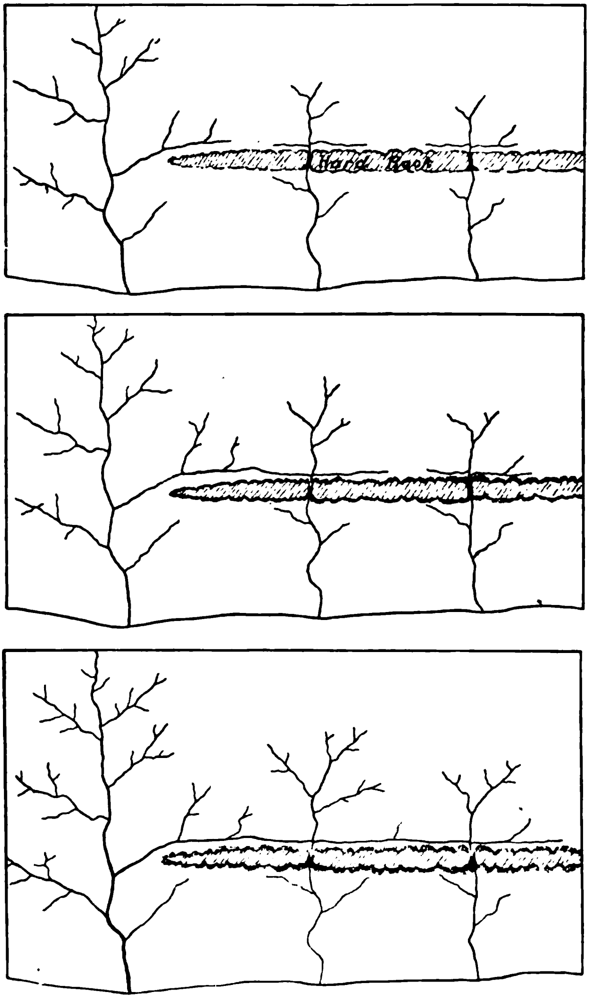


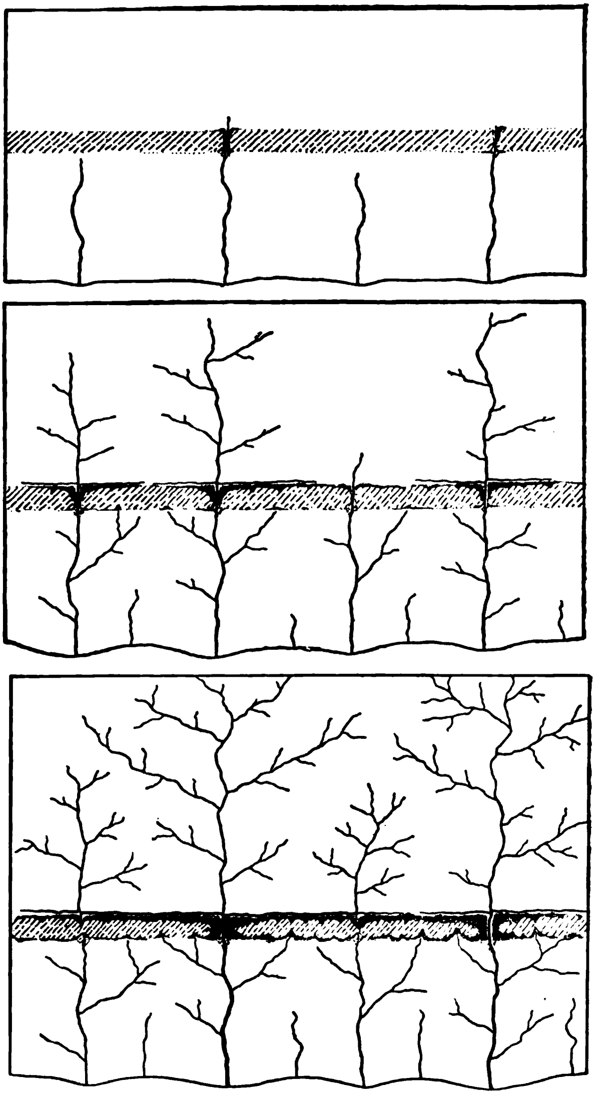
Fig. 124.—Matteo tepee, Wyo. Mass of igneous rock exposed by erosion, and preserved because of its superior resistance. (Detroit Photo. Co.)

Ridges and hills resulting from the unequal degradation of unequally resistant terranes are not equally prominent at all stages in an erosion cycle. In early youth the material surrounding the hard bodies of rock has not been removed; in early maturity considerable portions of their surroundings still remain about them; but in late maturity or early old age the outcropping masses of hard rock have been more perfectly isolated and are most conspicuous. Most of the even-crested ridges of the Appalachian system, as well as many others which might be mentioned, became ridges in this way. In the final stages of an erosion cycle the ridges of hard rock are themselves brought low. Isolated remnants of hard rock which remain distinctly



FIGS. 125–27.—Diagrams illustrating piracy, where the stream which does not flow over rock of superior hardness captures those which do. Fig. 126 represents a further development of the drainage shown in Fig. 125, and Fig. 127 represents a still later stage.

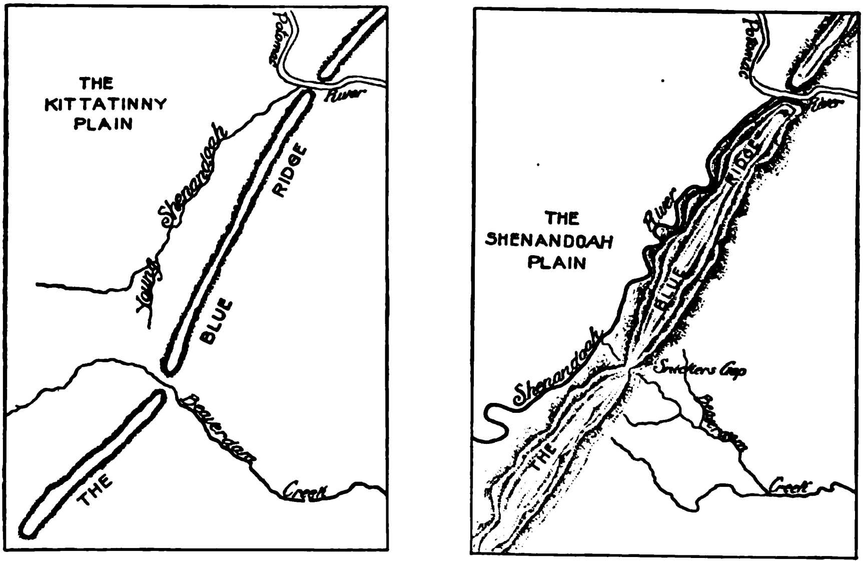
above their surroundings in the late stages of an erosion cycle (Fig. 124) are known as *Monadnocks*, the name being derived from Mount Monadnock, N. H., an elevation of this sort developed in a cycle antedating the present.



Figs. 128–30.—Diagrams to illustrate piracy, where the competing streams all cross a hard layer. The diagrams represent successive stages of development.

Adjustment of streams to rock structures.—Valleys (gullies) locate themselves at the outset without immediate regard to the hardness and softness of their beds. It is primarily the slope about the head of a gully which determines its line of growth, though relative hardness often determines the details of slope, even in the early stages of an erosion cycle. Once established, streams tend to hold their courses, even if this involves the crossing of resistant layers.

While a region where more and less resistant layers of rock come to the surface is in a youthful stage of erosion, some of the valleys (and therefore the streams) are likely to be located on the less resistant rock, some on the more resistant, and some partly on the one and partly on the other. The streams on the weaker rock will deepen their valleys more rapidly than the others, and those which flow across stronger and weaker rocks alternately will deepen their valleys more rapidly than those which run on hard rock all the time. The former conclusion is self-evident. The latter appears from the fact that rapids will be likely to develop at the crossing of each hard layer, thus accelerating erosion at those points. Such a stream therefore not only has less hard rock to erode than one which flows on resistant rock all the time, but it erodes that which it does cross much faster.



Figs. 131, 132.—The capture of the head of Beaverdam Creek by the Shenandoah Va.-W. Va. (After Willis.)

Streams which do not cross hard layers therefore have an advantage over those which do, and the tributaries to such streams, since they join deeper mains, have an advantage over the tributaries to the others. The valleys of the former may lengthen until their heads reach the latter, and capture their streams. This sequence of events is illustrated in the accompanying diagrams (Figs. 125–27). Even where several streams cross the same resistant bed, piracy is likely to take place among them, for some are sure to deepen their valleys faster than others, because of inequalities of volume, load, or hardness. This is illustrated by Figs. 128–30. An actual case is shown in Figs. 131, 132. Though piracy may take place when streams do not flow over rock of unequal hardness (p. 103), it is much more common where unequal resistance of the rock puts one stream at a disadvantage as compared with another.

The changes in the courses of streams, by means of which they come to sustain definite and stable relations to the rock structure beneath, are known as processes of *adjustment.*[1] Since streams and valleys adjust themselves to other conditions as well, this phase of adjustment may be called *structural adjustment.* Structural adjustment is not uncommon among rivers flowing over strata which are vertical or highly inclined, since in these positions the hard and soft strata are most likely to come to the surface in frequent alternation. The smaller streams suffer capture and adjustment first, since, as a rule, they have shallower valleys. It often happens that main streams, because of their deeper valleys, hold courses not in adjustment with structure (the Delaware, the Susquehanna, etc.), while tributary streams are captured, diverted, and adjusted. The capture of a tributary, however, leads both to the diminution of its main and to the increase of its captor, and the weakened stream may ultimately fall a prey to the one which is strengthened.

The processes of adjustment go on until the streams flow as much as possible on the weaker beds, and as little as possible on the stronger, when adjustment is complete. This amounts to the same thing as saying that the outcrops of the hard layers tend to become divides. In many cases an area is so situated that there is no escape for its drainage except across resistant rock. In this case its drainage is completely adjusted when as few streams as possible cross the resistant rock, and these by the shortest routes.

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[1] See Campbell, Jour. Geol., Vol. IV, pp. 567, 657.

Adjustment has been carried to a high degree of perfection in most parts of the Appalachian system. Here, as in all other mountains

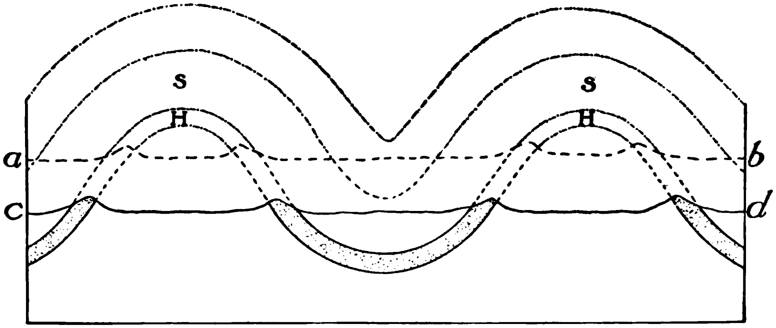

Fig. 133.—Diagram showing the outcrops of hard layers on the flanks of a truncated fold. *cd* represents the present surface; dotted lines above, earlier surfaces.

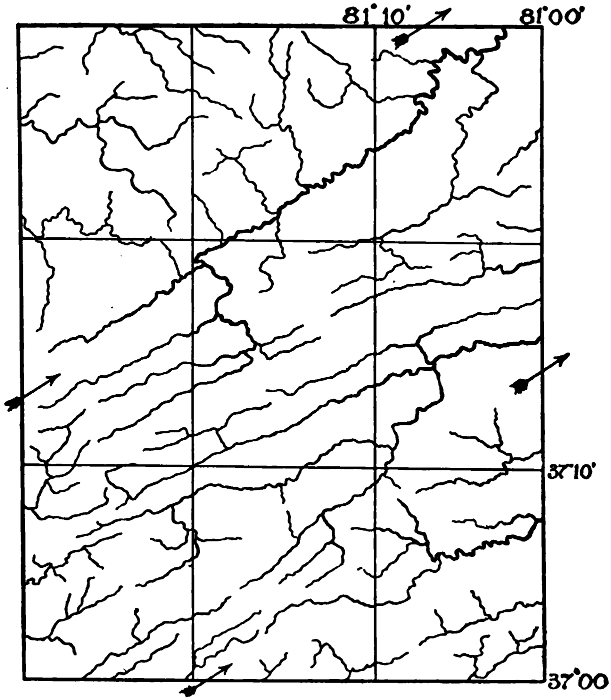

Fig. 134.—Example of adjusted drainage in a region of folded rocks, Va.-W. Va.

of similar structure, strata of unequal hardness were folded into ridges. In this case, the folds have been truncated by erosion, exposing the more and the less resistant beds (*H* and *S* respectively) in alternate belts along the flanks of the truncated folds (*ab* and *cd*, Fig. 133). The streams, especially the lesser ones, now flow along the strike of the

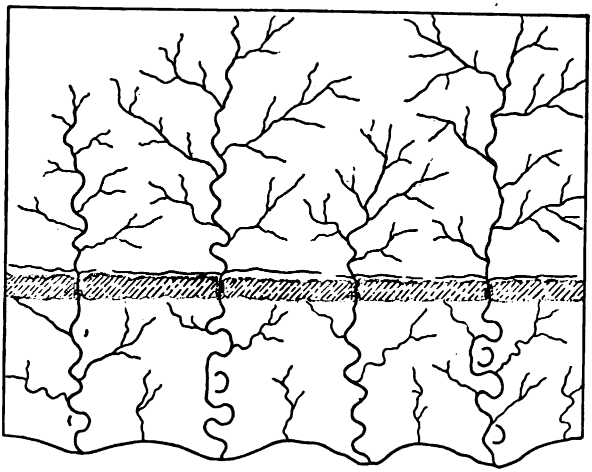

FIG. 135.—Diagram to illustrate readjustment of drainage, as base-level is approached.

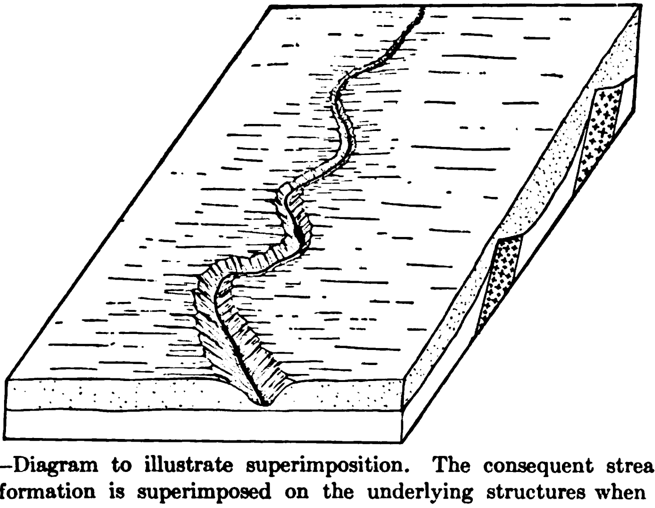

FIG. 136.—Diagram to illustrate superimposition. The consequent stream on the upper formation is superimposed on the underlying structures when the upper bed has been cut through.

softer beds much more commonly than elsewhere, and where they cross the hard layers it is usually at right angles to the strike. This is shown in Fig. 134, where the arrows indicate the direction of strike.

In the history of these rivers, however, a factor is involved which has not yet been considered, and these streams will be referred to later.

As base-level is approached, the outcrops of hard rock are brought low. When they have been reduced to the level of their surroundings, the streams may flow without regard to the resistance of the rock beneath, for downward cutting has ceased. As this stage of erosion is approached, a readjustment of the drainage may take place, and the waters which had taken long and circuitous courses to avoid hard rock, may change their courses to more direct ones (compare Figs. 130 and 135). Adjustment is, therefore, a relative term, and streams which are adjusted at one stage of erosion, are not necessarily adjusted at another.

It sometimes happens that rocks of unequal resistance are covered by beds of uniform hardness. A consequent stream developed on the latter may find itself out of structural adjustment when it has cut its channel down to the level of the heterogeneous beds below. Such a stream is said to be *superimposed* (Fig. 136) on the underlying structure. Structural adjustment is likely to follow.

INFLUENCE OF JOINTS AND FOLDS.

Joints.—Various structural features of rock other than hardness influence its erosion. Apart from the stratification planes, most rock formations are affected by joints or fissures. The joints are often, but not always, nearly vertical. Two sets are generally present, and sometimes more. If but two, they usually meet at a large angle; if more than two, two are likely to be nearly perpendicular to each other, while the third and fourth sets have such directions as to cut the others at large angles. These joints allow the ingress of water, roots, etc., which help to weather and disrupt rocks. Occasionally there is notable sag of the beds of rock along joint planes, but this effect is usually superficial only (Fig. 137). Where the jointage planes are frequent and open, the columns bounded by them sometimes topple over on cliff faces, either by undercutting, or by the wedge-work of roots or ice.

The effect of joints on erosion may often be seen along a stream which flows in a rock gorge. In such situations, the outlines of the banks are sometimes angular, and sometimes crenate (Fig. 138), the reëntrants being located at the joints. By working into and widening joints, running water sometimes isolates masses of rock as islands (Fig. 139). In a region free from mantle rock, or where the mantle rock is meagre, joints often determine the courses of valleys by directing the course of surface drainage. This is shown in many parts of the arid west. In regions where the rocks are notably faulted, the courses of the streams are sometimes controlled by the courses of the fault planes. This is the case, for example, in central Washington.[1]

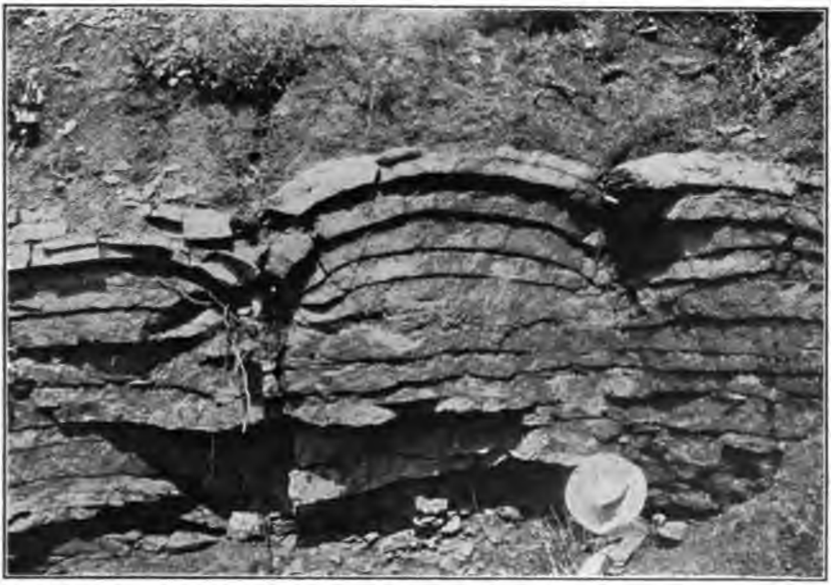

Fig. 137.—Shows the sagging of beds along joints. The disturbance does not extend far below the surface. Cook's quarry (Niagara limestone) near La Salle, Niagara Co., N. Y. (Gilbert, U. S. Geol. Surv.)

The jointing of rocks often shows itself distinctly in the weathered faces of cliffs (Figs. 140 and 141), especially in arid and semi-arid regions, or where the slope is too steep for the accumulation of soil and rock-waste on its surface.

If a stream flowing over jointed rock has falls, the conditions are sometimes afforded for the development of an exceptional and striking scenic feature. If above Niagara Falls, for example, there were

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[1] Russell. Rivers of North America, p. 280. The influence of joints on drainage is further discussed by Hobbs, Jour. Geol., Vol. IX, p. 469.

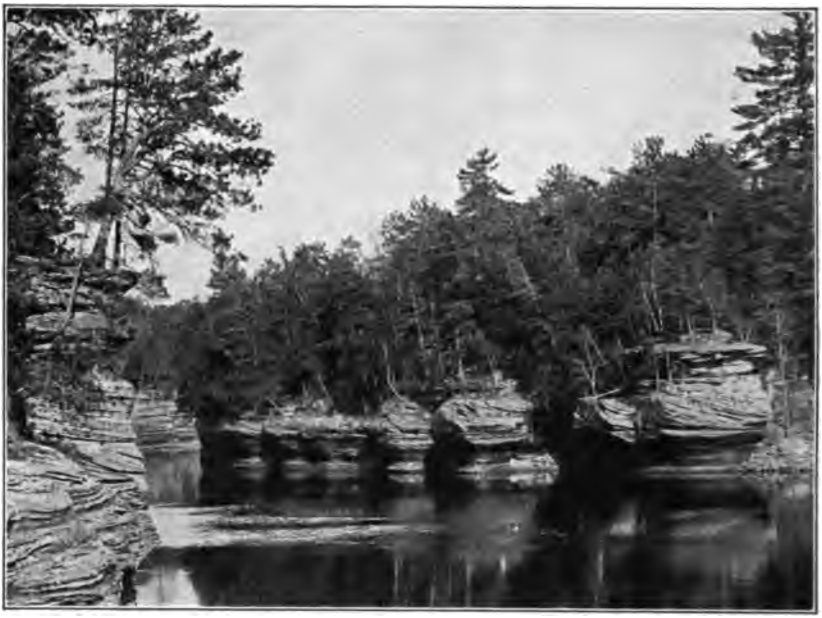

Fig. 138.—Figure showing crenate river bank, the reëntrants being determined by joints. Dells of the Wisconsin River, near Kilbourn, Wis. (Atwood.)

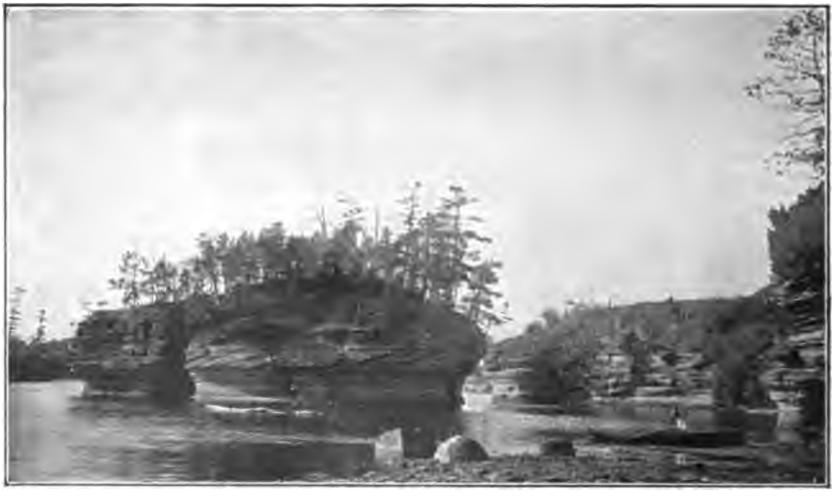

Fig. 139.—Lone Rock. An island isolated by the notable widening of a series of joints. The joints in the rock of the island have themselves been so widened that a rowboat may be taken through it in two directions. Lower Dells of the Wisconsin. (Meyers.)

an open joint in the bed of the stream (as at *b*, Fig. 142), some portion of the water would descend through it. After reaching a lower level it might find or make a passage through the rock to the river below the falls. If even a little water took such a course, the flow would enlarge its channel, making a passageway between the joint through which the water descended and the valley below the falls (*bcde*, Fig. 142). This passageway might become large enough to accommodate all the water of the river. In this case, the entire fall would be transferred from the position which it previously occupied (*f*) to the position of the enlarged joint (*b*). The fall would then recede. The underground channel between the old falls and the new would be bridged by rock (*bf''* and *f'''*, Fig. 143), making a *natural bridge*. The natural bridge near Lexington, Va. (Fig. 144), almost 200 feet above the stream which flows beneath it, is believed to have been developed

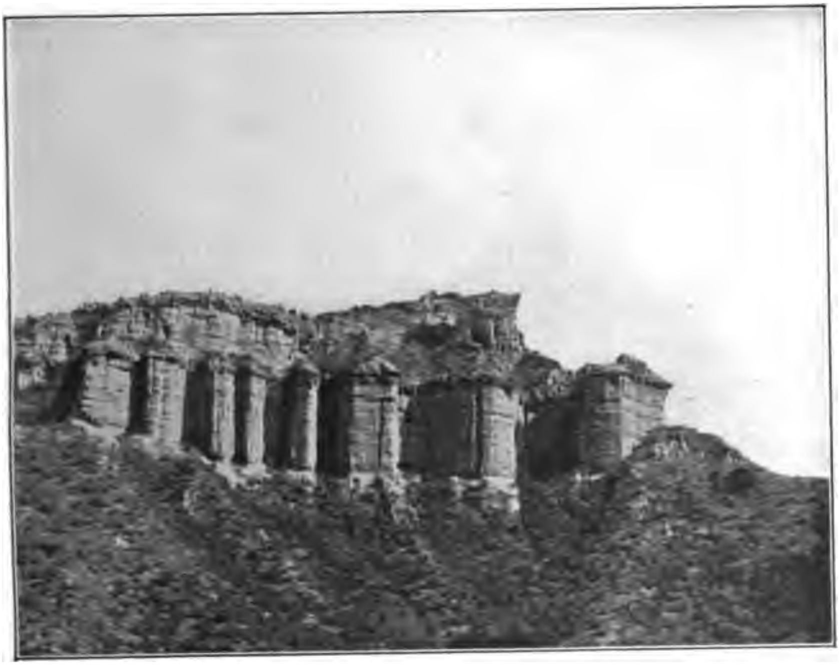

Fig. 140.—Effect of columnar structure on weathering. Material unconsolidated. Spur of south end of Sheep Mountain. (Lippincott, U. S. Geol. Surv.)

in this way. A similar bridge is now in process of development in Two Medicine River in northwestern Montana (Fig. 145). Once in existence, a natural bridge will slowly weather away.

It is not to be understood that all natural bridges have had this history. They are sometimes developed from underground caves when parts of their roofs are destroyed, as well as in various other ways.

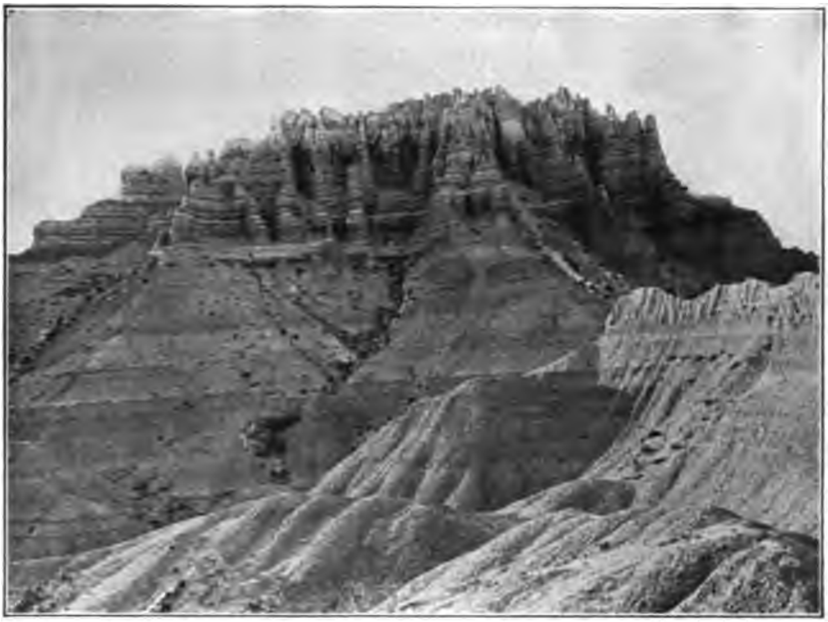

FIG. 141.—Effect of columnar structure on weathering. Big Bad Lands, S. D. (Darton, U. S. Geol. Surv.)

Folds.—The erosion of folded strata (anticlines and synclines) leads to the development of distinctive topographic features. So soon as a fold begins to be lifted, it is, by reason of its position, subject to more rapid erosion than its surroundings. For the same reason the crest of the fold is likely to be degraded more rapidly than its lower slopes, and must suffer more degradation before it is brought to base-level. Folds are usually composed of beds of unequal resistance, and as the degradation of a fold proceeds, successive layers are worn from the top, and the alternating hard and soft layers composing it are exposed. So soon as this is accomplished, adjustment of the streams is likely to begin, and the watercourses, and later the valley plains, come to be located on the outcrops of the less resistant layers, while the outcrops of the harder beds become ridges.

If the axis of an eroded anticline were horizontal, a given hard layer, the arch of which has been cut off, would, after erosion, outcrop on both sides of the axis. When the topography was mature these outcrops would constitute parallel ridges, or parallel lines of hills; when the region had been base-leveled, the outcrops would be in parallel belts, though no longer ridges or hills. The lower the plane of truncation, the farther apart would the outcrops be in the anticline, and the nearer together in the syncline (compare *ab* and *cd*, Fig. 133).

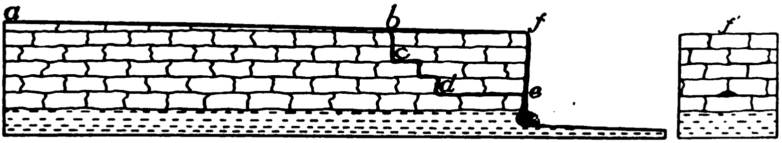

Fig. 142.—A natural bridge in process of development; longitudinal section at the left; transverse section, looking toward *e*, at the right.

If, on the other hand, the axis of the anticline or syncline to be eroded was not horizontal, that is, if it *plunged*, the topographic result would be somewhat different. Suppose a plunging anticline to be truncated at base-level. If either end of the fold plunged below the plane of truncation, the outcrops of a given layer on opposite sides of the axis would converge in the direction of plunge, and come together at the end. At a stage of erosion antedating planation (say late maturity) there would have been a ridge, or a succession of hills, in the position corresponding to the outcrop of a hard layer, with a canoe-shaped valley within. If two hard layers were involved, instead of one, there would be two encircling ridges, with a curved valley between them, and a canoe-shaped valley within the innermost (Fig. 146). If

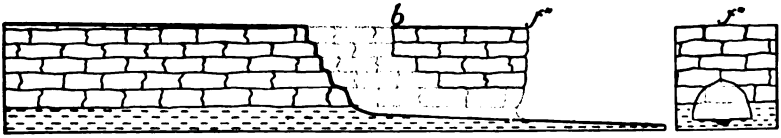

Fig. 143.—The same as Fig. 142 at a later stage of development.

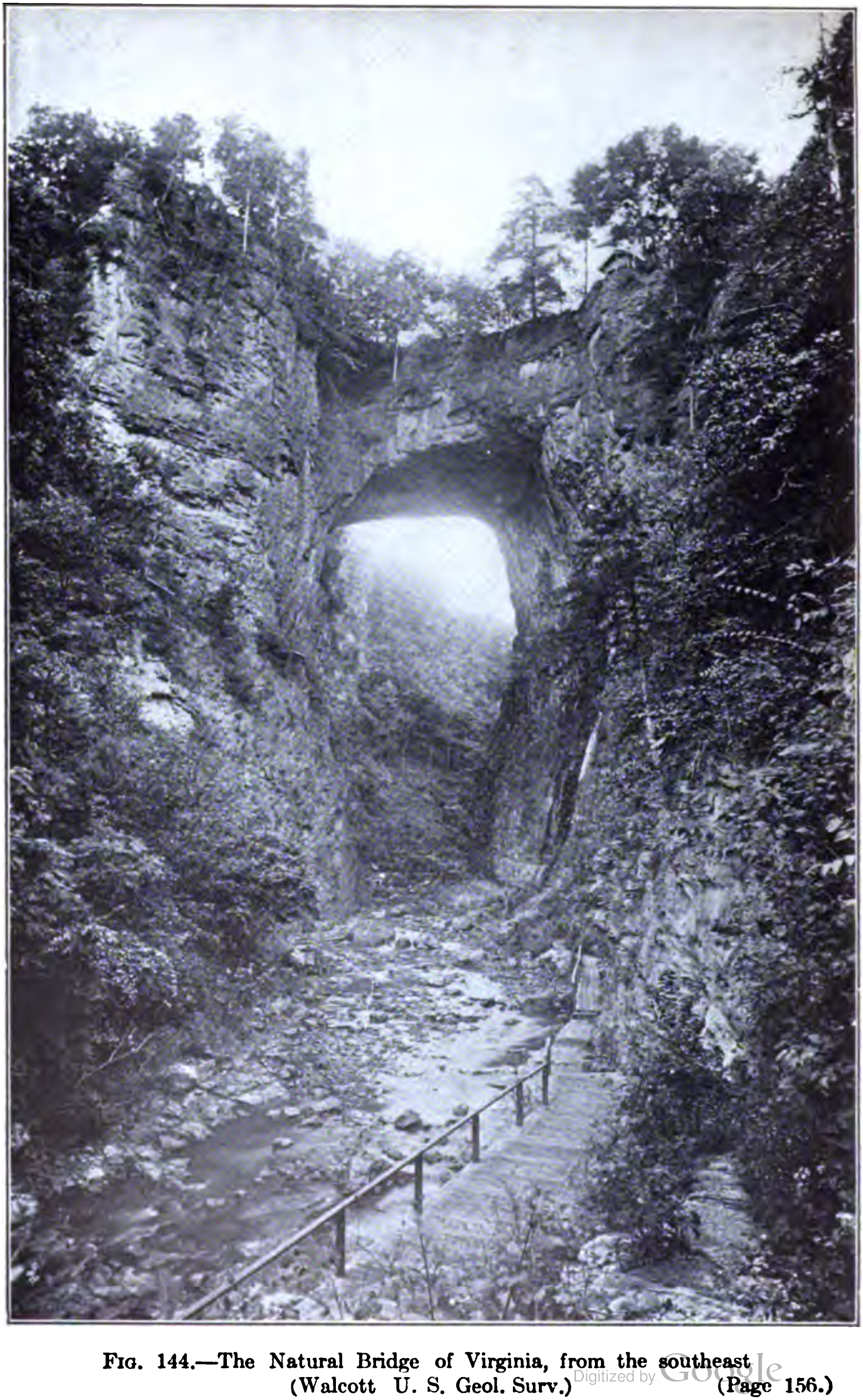

FIG. 144.—The Natural Bridge of Virginia, from the southeast (Walcott U. S. Geol. Surv.)

(Page 156.)

the anticline plunged both ways, the valley enclosed by the hard-layer ridge would be canoe-shaped at both ends (Fig. 147). In such a case there would be likely to be a low gap (water-gap) in the rim of the valley through which the drainage which degraded the surface escaped, but there would be likely to be but one, for if two or more streams had drained the area of the valley at an early stage of erosion, one would be likely to have captured the others (see p. 138) before late maturity. A succession of doubly-plunging anticlines and synclines might give rise to a very complex series of ridges and valleys. Illustrations of the above phenomena are found at various points in the Appalachian Mountains, especially in eastern Pennsylvania.[1]

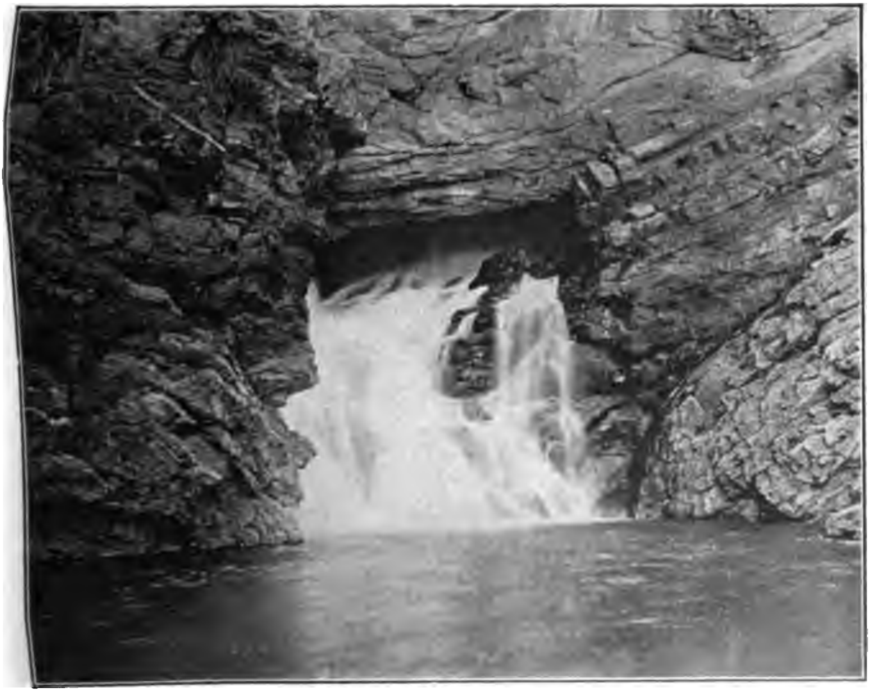

FIG. 145.—A natural bridge in development. Two Medicine River, Mont. Corresponds to the stage represented by Fig. 142, and the view corresponds to that shown diagrammatically at the right-hand end of the figure. (Whitney.)

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[1] See Willis. The Northern Appalachians, in Physiography of the United States.

In the structural adjustment which goes with the erosion of folds, it often happens that the valleys come to be located on the anticlines, while the outcrops of the hard layers on the flanks of the anticlines, or even in the original synclines, become the mountains. The ad-

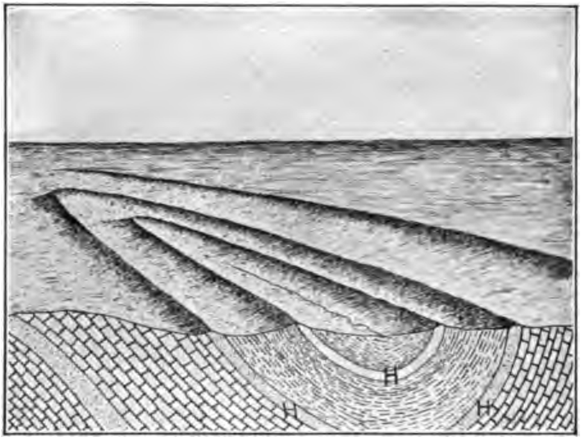


Fig. 146.—A canoe-shaped valley bordered by a ridge formed by the outcrop of a hard layer in a plunging syncline. The ridge bounding the canoe-valley is separated from an outer ridge by a curved valley underlain by relatively weak rock. (After Willis.)

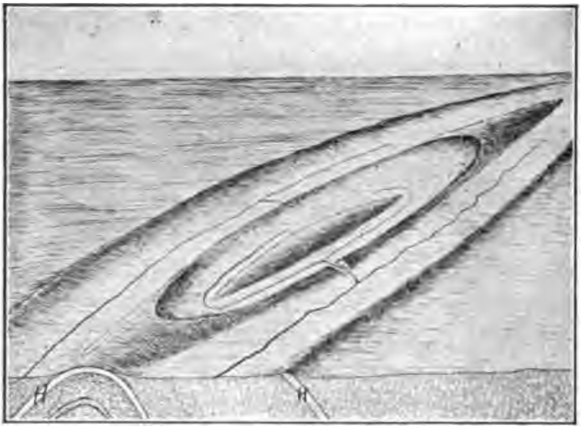
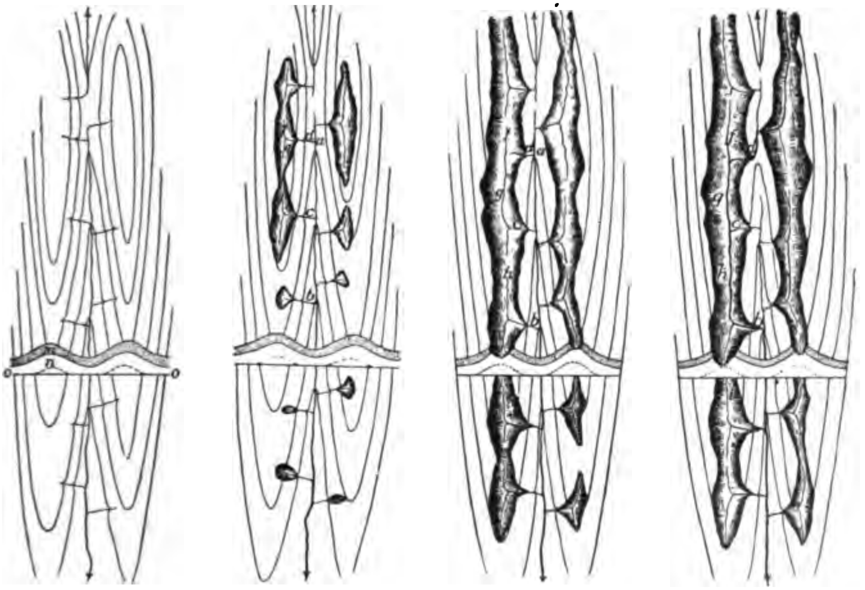


Fig. 147.—A diagram to illustrate the effects of erosion on a doubly-plunging anticline made up of beds of unequal hardness.

justments by which valleys come to be located on anticlines are somewhat as follows:[1] Fig. 148 represents two doubly-plunging anticlines with a syncline between, the relative elevations being shown by contour lines. At the outset, the drainage of such a region must have followed the structural valley, and its initial course, consequent on the slope, must have been down the axial trough. Drainage from the anticlines into the synclines would have promptly developed valleys, and the valleys would soon have acquired streams.



Figs. 148–51.—Diagrams to illustrate the shifting of rivers from a synclinal to an anticlinal position. (After Davis.)

The anticlines and synclines under consideration are assumed to have a thick hard layer at the surface, and softer beds below. This is shown in the cross-section introduced in the figure, the upper hard stratum (*m*) being indicated by the dots, while the softer one (*n*) is white. The line *oo* represents base-level, which is below the hard layer both in the syncline and anticline, but much farther below in the latter position than in the former. Because of their higher gradients, and because of the greater fracturing to which the region they drain was presumably subject at the time of folding, the tributary streams might cut through the hard layer sooner than the main stream which they join. This done, they would enlarge their valleys rapidly in the softer rock beneath, and secondary tributaries would be developed (Fig. 149). When the condition of things represented in Fig. 149 is reached, the streams *c* and *d*, tributary to the synclinal stream, come into competition. The former has the advantage over the latter, because it joins the main stream at a lower level. Stream *c* will therefore be likely to capture *d*. The incipient stages of the capture are stealthy, and the later bold. At first the divide between their head waters is shifted northward inch by inch, because the gradient toward *g* is higher than that toward *e*. The capture of the head waters of *e* is as slow as the migration of the divide, until the divide reaches the point where *e* joins *f*. The stream *f* is then diverted promptly into the valley of *g*, and is at once led away to *c* (see Fig. 150). Strengthened by its increased volume, the stream *c* (Fig. 150) lowers its valley across the hard layer more rapidly than before, and so holds the advantage it has gained. Not only this, but the beheaded stream *d* (Fig. 150), because of its diminished volume, sinks its valley into the hard layer less rapidly than before, and its decrease in power also works to the advantage of the stream leading to *c*. The result is that the divide between *fg* and *d* does not remain constant, but is driven back step by step toward *a*.

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[1] This process of adjustment has been well described by Davis in The Rivers and Valleys of Pennsylvania, Natl. Geog. Mag., Vol. I, p. 211 et seq.

Similarly a tributary to the main stream at *b* (Fig. 150), may by means of its tributary *h*, capture the waters of *fg*, and lead them to the synclinal valley at *b* (compare Figs. 150 and 151). Deprived of its main source of supply (at *c*) the synclinal stream is greatly diminished above *b*, and cuts more and more slowly, while the stream *fgh* (Fig. 151), having greater volume and working mainly in softer rock, sinks its channel faster than the stream in the synclinal axis. Under these circumstances, the stream at *f* may cut its valley below the valley in the synclinal axis *a* (Fig. 150). In this event, the divide between *f* and *a* (Fig. 150) may be pushed back until the synclinal stream is beheaded at *a* and carried out of the syncline and over into the anticlinal valley (Fig. 151). Thus, the old anticlinal axis comes to be the course of the main stream. Similarly the stream entering the syncline at *b* (Fig. 151) might later be captured by *i*, thus lengthening its anticlinal course.

It is not to be understood that this sequence of events will take place in the degradation of every anticline, but the principles here set forth will always be operative. The result specified will be accomplished wherever hard and soft layers have the relations indicated in the diagrams; that is, where the stream in the syncline finds itself on a resistant layer as it approaches base-level, while at the same time the (original) tributary streams are working in softer beds. It is not to be understood, therefore, that streams migrate from synclines to anticlines for the sake of getting out of the former positions into the latter. If they shift their courses it is to find easier ones.

That these changes are not fanciful is shown by the fact that the adjustment described corresponds with that shown in many parts of the Appalachian Mountains, and in other mountains of similar structure.

If in a later stage of its history, the new main stream, *fh*, were to cut its bed down to a lower hard layer, while the original stream, *ab*, reached a softer bed beneath the hard one above, the latter would again have an advantage, and a new series of adjustments would be inaugurated which might result in re-establishing the main stream in its original synclinal position.

EFFECT OF CHANGES OF LEVEL.

Rise.—If after being base-leveled, or notably reduced by erosion, a region is uplifted so as to increase the gradients and therefore the velocities of the streams which drain it, the streams are said to be *rejuvenated*, and a new cycle of erosion is begun. If the rise of the area were equal everywhere, while the coast line remained constant in position, there would be an immediate increase in velocity only at the debouchures of the streams flowing directly into the sea. At the debouchures of such streams there would be rapids or falls. Each

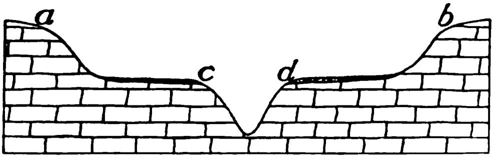

Fig. 152.—Cross-section of a wide valley, *ab*, in the bottom of which a younger valley, *cd*, has been excavated as the result of uplift.

rapids or falls would promptly recede, and with the recession, the acceleration of velocity resulting from the uplift would be felt farther and farther up-stream, and ultimately to its source. The rejuvenated streams would cut new valleys in the bottoms of their old ones (Figs. 152 and 153). The new valleys would begin where the increase in velocity was first felt, and they would be lengthened by head erosion just as valleys of the first cycle were lengthened.

When the head of the new part of a valley of a rejuvenated stream recedes past the mouth of a tributary adjusted[1] to the gradient of the main stream before rejuvenation, the velocity of the tributary is

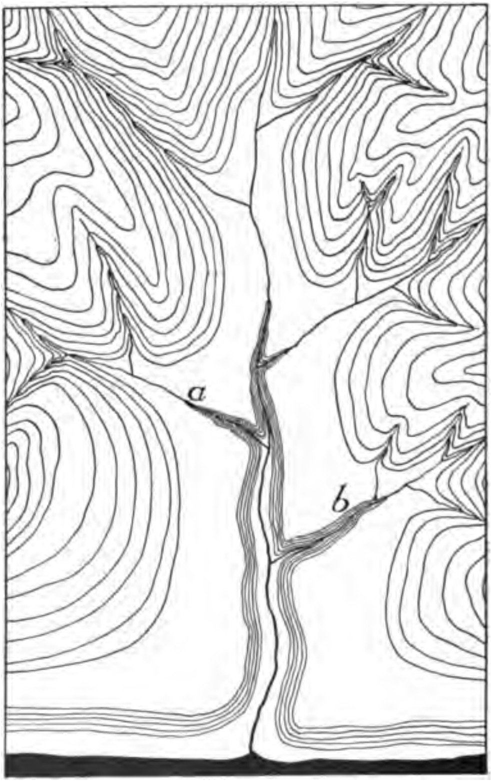

Fig. 153.—Diagram to illustrate in ground plan an ideal case of rejuvenation as the result of uplift.

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[1] This sort of adjustment may be called *topographic adjustment*. A tributary is in topographic adjustment when its gradient is harmonious with that of its main.

PLATE XIII.

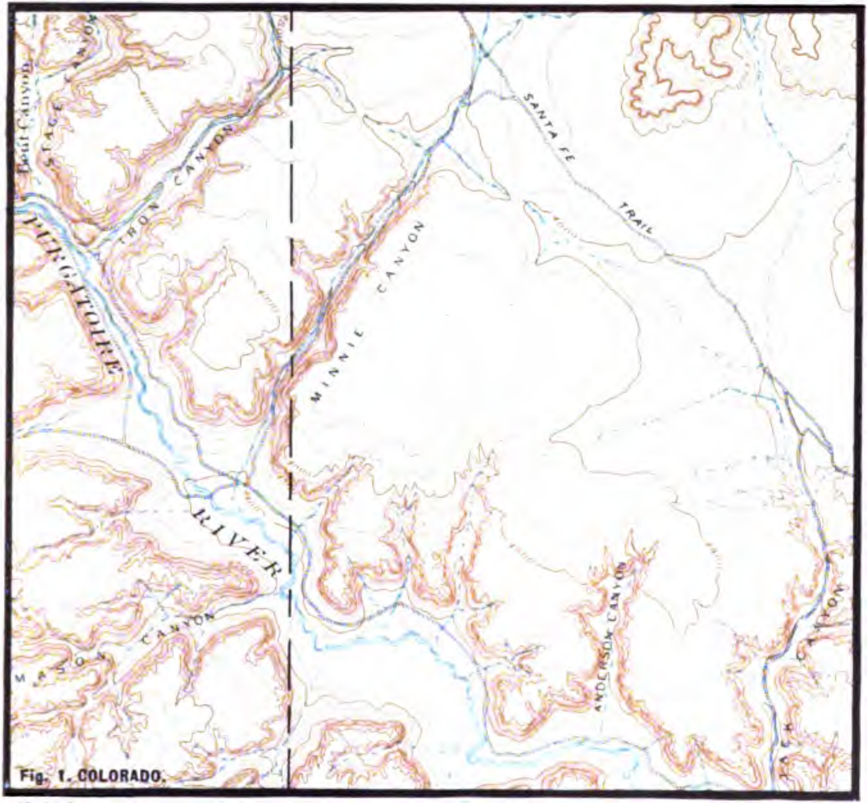

Fig. 1. COLORADO.

U. S. Geol. Surv.

Scale, 2+ miles per inch.

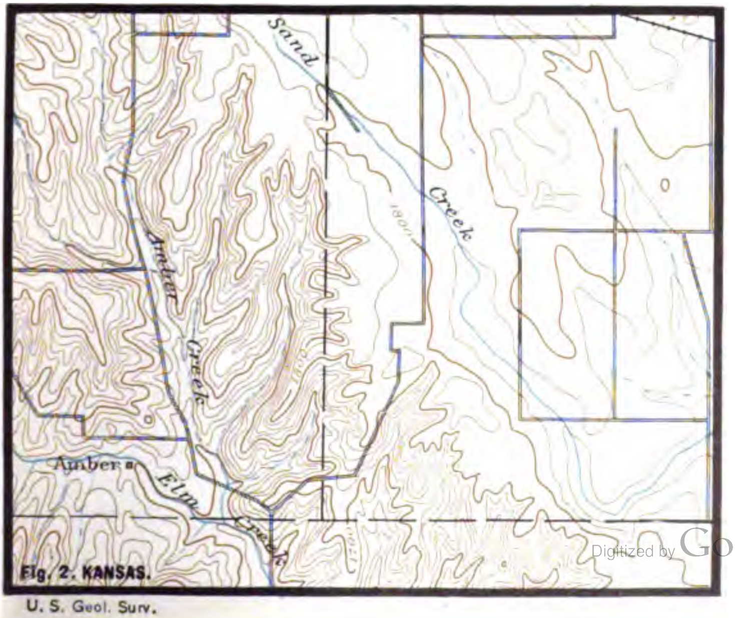

Fig. 2. KANSAS.

U. S. Geol. Surv.

PLATE XIV.

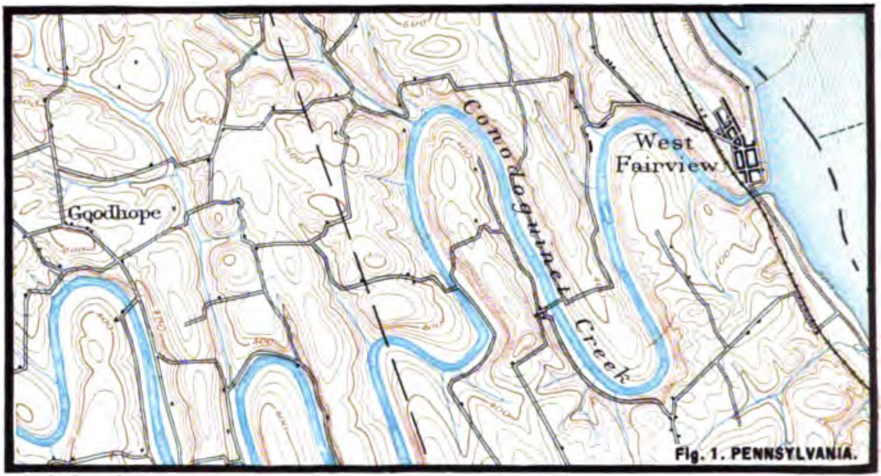

Fig. 1. PENNSYLVANIA.

U. S. Geol. Surv.

Scale, 1+ mile per inch.

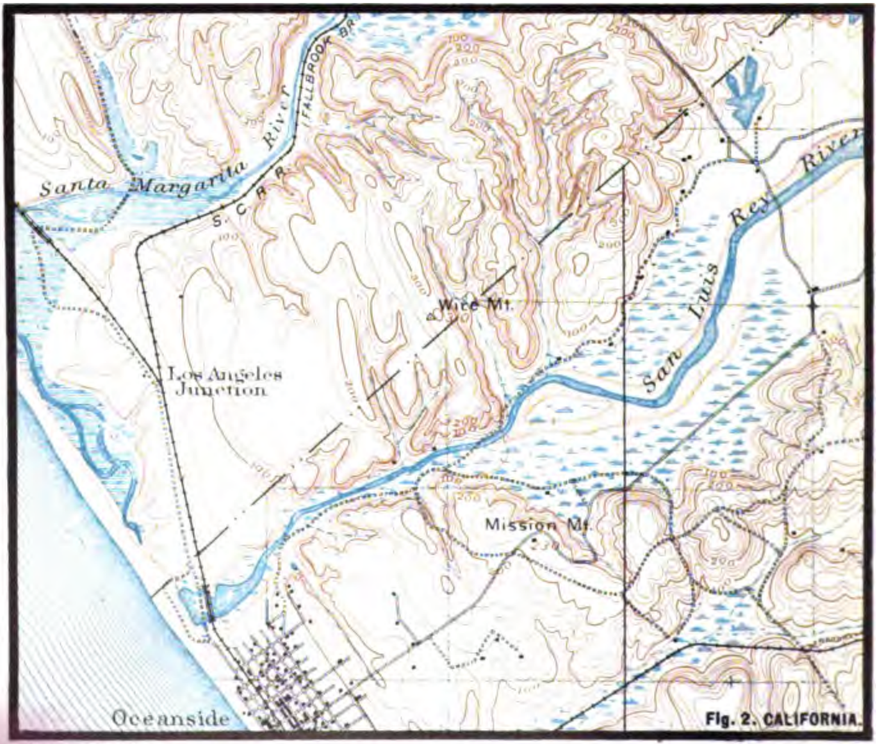

Fig. 2. CALIFORNIA.

U. S. Geol. Surv.

Scale, 1+ mile per inch.

accelerated at its debouchure, and it begins to excavate a new valley in the bottom of its old one. The new valley commences at the lower end of the old one, and develops headward (*a* and *b*, Fig. 153). Good illustrations are furnished by the streams in the west central part of New Jersey. The Delaware has here a sharply defined valley, and its tributaries are essentially as deep as their main at the point of junction. Above this point they have high gradients for a short distance (three to six miles), beyond which they wind sluggishly in wide valleys with low gradients across a relatively high plateau. Their profiles are illustrated by Fig. 154. The flat, though high, surface in which their upper courses lie, appears to have been nearly base-leveled in an earlier cycle, and then to have been elevated. The date of the elevation is fixed, in terms of erosion, by the time necessary for the excavation of the Delaware gorge, and the narrow gorges along the lower courses of its tributaries. It was so recent that the effects of rejuvenation, proceeding from the debouchures of the tributaries toward their heads, have not yet advanced far from the Delaware. Similar relations are found elsewhere (Fig. 1, Pl. XIII, s. c. Col.). Another peculiarity of rejuvenated drainage is shown in Fig. 2, Plate XIII (s. Kan.). Here Elm Creek flows at a level 200 feet below that of Sand Creek, 4 miles distant. The valley of the former appears to have entered upon a new cycle as the result of uplift, while that of the latter, *in the area shown on the map*, is still unrejuvenated. Farther down-stream, the valley of Sand Creek shows signs of rejuvenation. It may be noted that a tributary of Amber Creek has good opportunity to capture Sand Creek, for the latter flows about 25 miles before reaching the level of Amber Creek at its junction with Elm Creek.

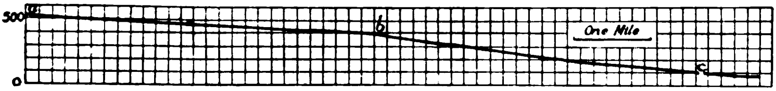

Fig. 154.—Profile of a rejuvenated stream. The Lockatong River (N. J.) to the head of Mud Run.

Should the lower end of a tributary valley fail to be degraded as fast as the valley of the main at the point of junction, the tributary is out of *topographic adjustment* with its main. Falls or rapids may result. When the lower end of a tributary valley is distinctly above the level of its main, the former is called a *hanging valley*. Hanging valleys developed by stream erosion alone are not common except just after the recession of a falls past the mouth of a tributary. Hanging valleys, as well as the characters and relations illustrated by Figs. 152–154 are criteria of rejuvenation, but they must be applied with discretion. Such profiles, for example, as that shown in Fig. 154 may be developed when the rock of a stream's bed is unequally resistant, and hanging valleys are generally a result of glaciation (see Chapter V).

Rejuvenated streams sometimes inherit certain peculiarities from their aged ancestors. Thus a rejuvenated stream may intrench the meanders possessed by the old stream which preceded (see Fig. 1, Pl. XIV, near Harrisburg, Pa.), and intrenched meanders are one of the marks of rejuvenated streams. They are not uncommon in the Appalachian Mountain regions, and are known in other parts of the world. The Seine and the Moselle furnish further illustrations.[1]

The history of the new cycle of erosion inaugurated by the uplift would differ from that of the preceding cycle in that the new one would begin with a drainage system already developed. Other things being equal, therefore, the reduction of the land would proceed more rapidly in a subsequent cycle than in the first.

The recognition of different cycles of erosion, separated by uplifts, is often easy. The principles involved are illustrated by Fig. 155 which represents an ideal profile of considerable length (say 50 miles). The points *a*, *a′*, and *a″* reach a common level. Below them there are areas *b*, *b′*, and *b″* which have a nearly common elevation, below which are the sharp valleys *d*, *d′*, and *d″*. The points *a*, *a′*, and *a″* represent the cross-sections of ridges formed by the outcrops of layers of hard rock. If the crests of the ridges are level, the points *a*, *a′*, and *a″* must represent remnants of an old base-level, *since at no time after a ridge of hard rock becomes deeply notched does it acquire an even crest, until*

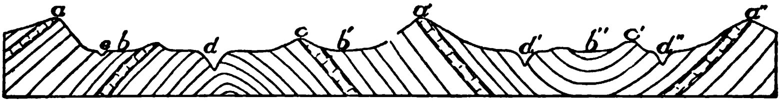

Fig. 155.—Diagram to illustrate cycles of erosion where the beds are tilted.

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[1] Davis. The Seine, the Meuse and the Moselle. Nat'l Geog. Mag., Vol. VII, pp. 181–202, and 228–238. An article which throws much light on the behavior of rivers.

it is base-leveled.[1] At all earlier stages its crest is uneven. After the cycle represented by the remnants *a*, *a′*, and *a″* was completed, the region suffered uplift. A new cycle represented by the plain *b*, *b′*, and *b″* was well advanced, though not completed, when the region was again elevated, and the rejuvenated streams began to cut their valleys *d*, *d′*, and *d″* in the plain of the previous incomplete cycle. The elevations, *c* and *c′* (intermediate in elevation between *a*, *a′*, and *a″*, and *b*, *b′*, and *b″*) may represent either remnants of the first base-level plain which were lowered, but not obliterated, while the plane *b*, *b′*, *b″* was developing; or they may represent a cycle intermediate between that during which *a*, *a′*, *a″* and *b*, *b′*, *b″* were developed. If the intermediate elevations (*c*, *c′*) have a common height and level crests, the presumption would be in favor of the latter interpretation. If they be numerous and of varying heights, as is possible, they may in the field obscure the planes (*a*, *a′*, *a″* and *b*, *b′*, *b″*) developed in the different cycles, which, in the figure, are distinct.

If the strata involved be horizontal the determination of cycles is sometimes less easy. Thus in Fig. 156, it is not possible to say whether *a* and *a′* represent remnants of an old base-level, or whether they represent the original surface from which degradation started. So, too, the various benches below *a*, such as *b*, *b′*, and *b″* may readily be the result of the superior hardness of the beds at this level. For the determination of successive uplifts in the field it is necessary to consider areas of considerable size, and to eliminate the topographic effects of inequalities of hardness, and of certain other factors to be mentioned presently.

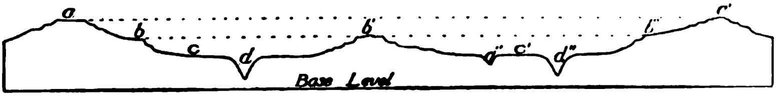

FIG. 156.—Diagram to illustrate cycles of erosion where the beds are horizontal.

The inequalities in the depths of the young valleys in Figs. 155 and 156 may be explained on the supposition that the deeper ones belong to main streams, and the shallower ones to tributaries. Such a valley as that shown at *e*, Fig. 155, suggests rejuvenation at this point; but farther up the stream which occupies this valley, rejuvenation might not be apparent. In this case, the main streams might be flowing in new valleys, *d*, *d'*, etc., while the heads of their tributaries are still flowing in the older valleys of the preceding cycle (compare Fig. 154 and Fig. 1, Pl. XIII).

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[1] Another view has been advocated by Tarr, Am. Geol. Vol. XXI, pp. 351–370.

It is by the application of the preceding principles that it is known that the Appalachian Mountains, after being folded, were reduced to a peneplain (p. 76), throughout their whole extent from the Hudson River to Alabama. The peneplain level is indicated by the level crests of the Appalachian ridges, shown in cross profile by the high points of Fig. 157. The system was then uplifted, and in the cycle of erosion which followed, broad plains were developed at a new and lower level, corresponding in a general way to the plains *b*, *b'*, and *b''* of Fig. 155. The plains were located, for the most part, where the less resistant strata come to the surface. Above them rose even-crested ridges, the outcrops of the resistant layers, which had been isolated by the degradation of the softer beds between. They constitute the present mountain ridges (the high points of Fig. 157). The evenness of their crests, testifying to the completeness of the first peneplanation, is shown in Fig. 158, which represents, diagrammatically, a longitudinal profile of an Appalachian Mountain ridge. The evenness of the crest is interrupted by (1) notches (*b*, *c*, etc., Fig. 158) cut by the streams in later cycles, and (2) by occasional elevations above the common level (monadnocks, *a*, *a'*, Fig. 158). The monadnocks are generally rather inconspicuous, but there is a notable group of them in North Carolina and Tennessee. Mount Mitchell and Roane Mountain are examples. When long distances are considered, the ridge crests depart somewhat from horizontality. This is believed to be due, in part at least, to deformations of the old peneplain during the uplift which inaugurated the second cycle of erosion.

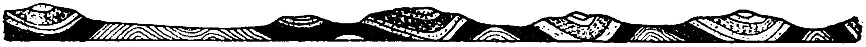

FIG. 157.—Cross-section of a portion of the Appalachian Mountains to illustrate the phenomena of erosion cycles. (After Rogers.)

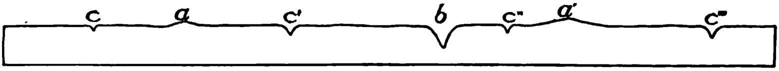

FIG. 158.—A diagrammatic longitudinal profile of an Appalachian Mountain ridge.

The extent to which the second cycle of erosion recorded in the present topography had proceeded before its interruption by uplift, is indicated by the extent of the valley plains (Fig. 157) below the mountain ridges. While these plains were being developed on the weak rocks, narrow valleys only (Fig. 158) were cut in the resistant rocks which now stood out as ridges. In Fig. 158 some of these valleys are shallow (*c*, *c'*, *c''*, etc.), and but one of them deep. The former may be either (1) the valleys of streams which crossed the hard layer at the beginning of the cycle, and which were diverted before their valleys became deep; or (2) they may represent the heads of valleys now working back into the ridges. The deep valley (*b*) represents the work of a stream which has held its course across the hard layer while the latter was being isolated as a mountain ridge (compare Figs. 131 and 132). Deep narrows of this sort are often called *water-gaps*. Similar valleys, whether shallow or deep, from which drainage has been diverted, are sometimes called *wind-gaps*. The second cycle of erosion, while still far from complete, was interrupted by uplift (relative or absolute), and a new cycle inaugurated. This event was so recent that the new (third) cycle has not yet advanced far.

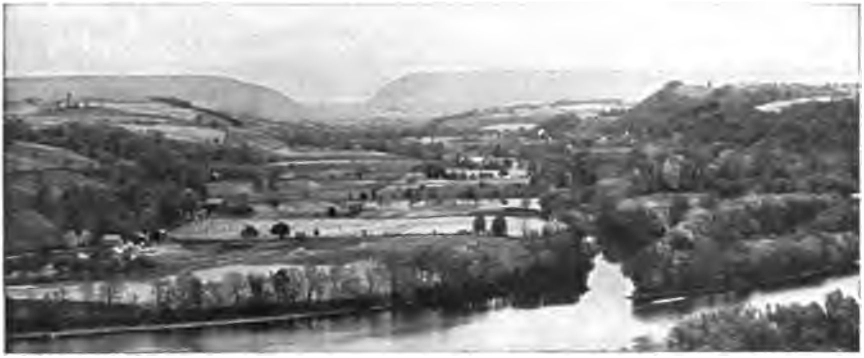

FIG. 159.—The Kittatinny Mountains and Delaware Water-Gap from Manunka Chunk. (N. J. Geol. Surv.)

Recently it has been urged that another cycle, intermediate between the first and second, is to be recognized.[1]

Some of the features just described are illustrated by Fig. 159. The even mountain crest in the background is the Kittatinny Mountain of New Jersey and its continuation in Pennsylvania. In common with other corresponding crests it represents the oldest recorded base-level (or peneplain) of the region. The great gap in the mountain is the Delaware Water-Gap. Below the mountain crest there is another plain, developed in a subsequent cycle of erosion, while the valley plain in the foreground represents the work of a still later cycle.

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[1] Campbell. Bull. Geol. Soc. of Am., Vol. XIV, p. 277.

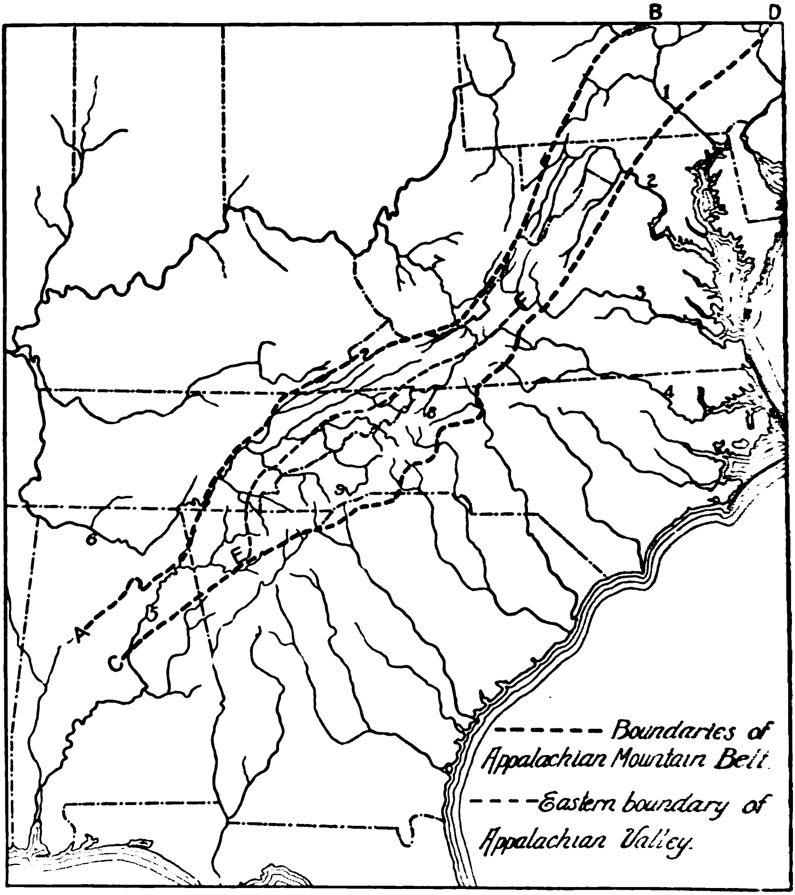
B
D
1
2
3
4
5
6
7
8
9
A
C
E
------ Boundaries of Appalachian Mountain Belt.
----Eastern boundary of Appalachian Valley.

FIG. 160.—Showing certain peculiarities of Appalachian drainage. 1 = the Susquehanna; 2 = the Potomac; 3 = the James; 4 = the Roanoke; 5 = the Coosa; 6 = the Tennessee; 7 = the Kanawha; 8 = head of New River; 9 = head of the French Broad

The oldest erosion plain of the Appalachian Mountains, the results of which are seen in the even-crested ridges so characteristic of the system, is sometimes called the *Kittatinny base-level.*[1] It was conpleted early in the Cretaceous period, and hence is sometimes known as the *Cretaceous base-level.* The next lower plain, imperfectly developed, has been called the Shenandoah Plain,[1] from the Shenandoah Valley where it is well seen (Fig. 132 and Fig. 2, Pl. XII). It is to be noted that the terms *base-level* and *peneplain* have both been used in connection with these old plains. *Graded plain* is equally applicable. The truth is that the topographic types represented by these three terms grade into one another. It may be questioned whether definitions should be insisted on which differentiate these types more sharply than Nature has.

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[1] Willis. Physiography of the United States. The Northern Appalachians.

Many of the peculiarities of the drainage of the Appalachian Mountain system are intimately connected with the history just outlined. Thus three great rivers, the Delaware, the Susquehanna, and the Potomac, have their sources west of the Appalachians proper, cross the system in apparent disregard of the structure, and flow into the Atlantic. The James and Roanoke head far to the west, although not beyond the mountain system, and flow eastward, while the New River (leading to the Kanawha) farther south, heads east of the mountain-folds, and flows northwestward across the alternating hard and soft beds of the whole Appalachian system, to the Ohio (Fig. 160). The French Broad, a tributary to the Tennessee, has a similar course. Such streams are clearly not in structural adjustment, and afford good opportunities for piracy. Their courses were apparently assumed during the time of the Kittatinny base-level, when the streams had so low a gradient as not to be affected by the structure (p. 150). Elevation rejuvenated them, and they have held their courses in succeeding cycles across beds of unequal resistance, though smaller streams have become somewhat thoroughly adjusted. Crustal deformations have also helped them to hold their courses, for the Cretaceous peneplain seems to have been tilted to the southeast at its northern end, and to the southwest at its southern, when the succeeding cycle began.

Streams which hold their early courses in spite of changes which have taken place since their courses were assumed are said to be *antecedent.* They antedate the crustal movements which, but for preexistent streams, would have given origin to a different arrangement of river courses. As a result of crustal movements, therefore, a consequent stream may become antecedent. Master streams are more likely to hold their courses, and therefore to become antecedent, than subordinate ones.

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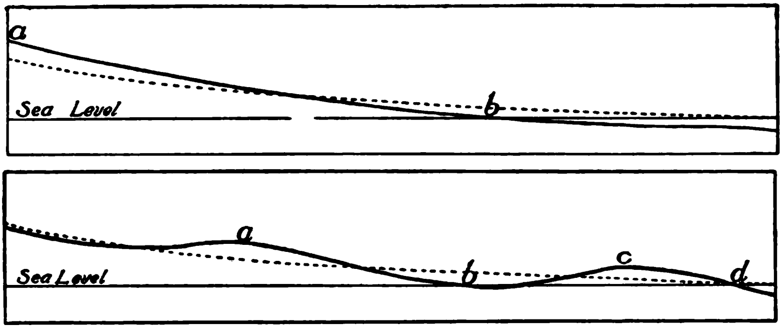
[1] Willis. Physiography of the United States. The Northern Appalachians.

The uplift of base-leveled beds, especially if the beds are tilted so as to bring layers of unequal resistance to the surface at frequent intervals, affords conditions favorable for extensive adjustment. The numerous wind-gaps in the mountain ridges, representing the abandoned courses of minor streams, and the less numerous water-gaps, which indicate the resistance of large streams to structural adjustment, are instructive witnesses of the extent to which adjustment has gone. So extensive has been the adjustment among the streams of the Appalachian Mountains that there is probably no considerable stream in the whole system which has not gained or lost through its own or its neighbors' piracy. The history of the rivers of the Appalachian Mountains has been further complicated by a considerable amount of warping during the periods of uplift.[1]



FIGS. 161, 162.—Diagrams to illustrate the effect of crustal warping on stream erosion. The dotted lines represent the profiles of the streams before deformation; the full lines, after. Erosion will be stimulated between *a* and *b* in each case, and between *c* and *d* in Fig. 162. Below *b*, Fig. 161, the stream will be drowned, and erosion therefore stopped. Erosion will also be stopped or retarded above *a*, between *b* and *c*, and below *d* in Fig. 162.

Sinking.—The land on which a river system is developed may be depressed relative to sea-level. In this case the sea would occupy the lower ends of valleys, converting them into bays and estuaries. A stream in this condition is said to be *drowned*. Of drowned rivers there are many examples along the Atlantic coast. Thus the St.

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[1] For excellent accounts of the rivers of the Appalachian Mountains see Davis, Rivers of Northern New Jersey, Nat'l Geog. Mag., Vol. II, pp. 81–110; and Rivers of Pennsylvania, op. cit., pp. 183–253; Willis, The Northern Appalachians, Physiography

Lawrence River is drowned up to Montreal, and the Hudson up to Albany. If the drowned portion of the latter valley were not so narrow, it would be a bay. Delaware and Chesapeake Bays, as well as many smaller ones, both north and south, are likewise the drowned ends of river valleys (see figures, Chapter VI). If all parts of a drainage basin sank equally, the velocities of the streams above the limit of drowning would not be changed, for the gradients would remain the same as before. The fact that a river's channel is below sea-level is not to be taken as proof that the valley is drowned. Thus the bottom of the channel of the Mississippi is as much as 100 feet below the level of the Gulf, some 20 miles above New Orleans.[1]

Differential movement. Warping.—Where a land surface on which a river system is established suffers warping, some parts going up and others down, the opposite movements being either absolute or relative, various phenomena would result. This may be illustrated by the accompanying diagrams (Figs. 161 and 162), where the profiles of the streams are represented as warped from the positions represented by the dotted lines, to the positions shown by the full lines. The velocity will be accelerated below the points of differential elevation (between *a* and *b*, Fig. 161, and between *a* and *b*, and *c* and *d*, Fig. 162), but checked above (above *a*, and between *b* and *c*, Fig. 162). Above an elevation which notably checks its flow, a stream is *ponded.* If the ponding is slight, a marsh may develop above the obstruction; if more considerable, a lake is formed. Lakes of this class are likely to be short-lived, since the ponded waters are likely to soon overflow and lower their outlet so as to drain the lake. The elevation which ponds the stream may be great enough and rapid enough so that the resulting lake finds an outlet by some course other than that originally followed by the stream. Where a stream holds its course across an uplift athwart its valley, either with or without ponding, it becomes an antecedent stream (see p. 169), since it has a course assumed before the latest deformation of the crust and in apparent disregard of present surface configuration. Thus the Columbia River holds

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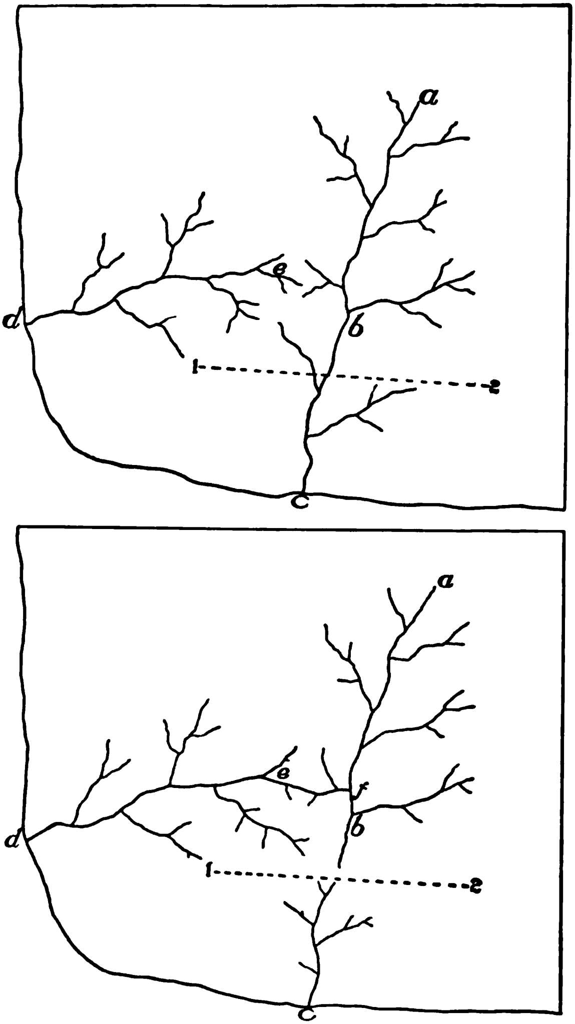
of the United States, pp. 169–202; Hayes, the Southern Appalachians, op. cit., pp. 305–336; Hayes and Campbell, The Geomorphology of the Southern Appalachians, Nat'l Geog. Mag., Vol. VI, pp. 63–126, and Hayes, Physiography of the Chattanooga District, 19th Ann. Rep. U. S. Geol. Surv., Pt. II, pp. 1–58.

[1] This is the case at Davis and Lone Star. Capt Howell, Miss. Riv. Commission.

its antecedent course across areas which have been uplifted (differentially) hundreds and even thousands of feet.[1] Some of the striking scenic features of this noble valley are the result of these changes in the country through which it flows. A lesser stream would have been



Figs. 163, 164.—Piracy stimulated by warping. **Uplift along axis 1–2.**

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[1] Russell. Rivers of North America, p. 279.

diverted, as many of its tributaries have been. Even its course across the Cascade ranges is believed to be antecedent.[1]

Another peculiarity of valleys and streams resulting from changes of level is illustrated in Fig. 2, Pl. XIV (southern California). The main valleys of this part of the coast were developed when the land stood considerably higher than now. Later the subsidence of the coast converted the lower ends of the valleys into bays or fiords. The bays were then transformed into lagoons by deposition. Subsequent rise of the land or depression of the sea allowed the drainage from the old lagoons to cut across the deposits which had converted the bays into lagoons. The result is an old, wide valley above, suggested by a young one below.

If the warpings were considerable, much more decisive changes in drainage would result. Suppose the drainage of a given region to be represented by the streams in Fig. 163. If there is uplift along the axis 1–2, that part of *ac* above the axis of uplift would be ponded, or at least have its velocity checked, while the flow of some of the tributaries of *d* would be accelerated, and might work back and capture the other stream (Fig. 164).

Crustal warping was one of the conditions under which the Tennessee achieved its present anomalous course, and its history[2] is illustrative of the complex changes which drainage suffers when warping affects the area where the rock structures are of unequal resistance. At the close of the Cretaceous cycle of erosion, when the Appalachian Mountains had been reduced to a peneplain, the waters falling in the area now drained by the upper course of the Tennessee flowed south-southwest to the Gulf in a stream (the Appalachian River, *a*, Fig. 165) the lower part of which had the general position of the Coosa and the Alabama.

To the west of the Appalachian River, shorter streams flowed west and southwest into the Mississippi embayment (Fig. 165) by courses which are not now definitely known. The succeeding cycle of erosion was inaugurated by uplift and deformation of the peneplain. The axis

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[1] Russell. Rivers of North America, p. 279.

[2] Hayes. Physiography of the Chattanooga District, 19th Ann. Rep., U. S. Geol. Surv., Pt. II, pp. 9–58. See, also, Hayes and Campbell, Geomorphology of the Southern Appalachians, Nat'l Geog. Mag., Vol. VI, pp. 63–126.

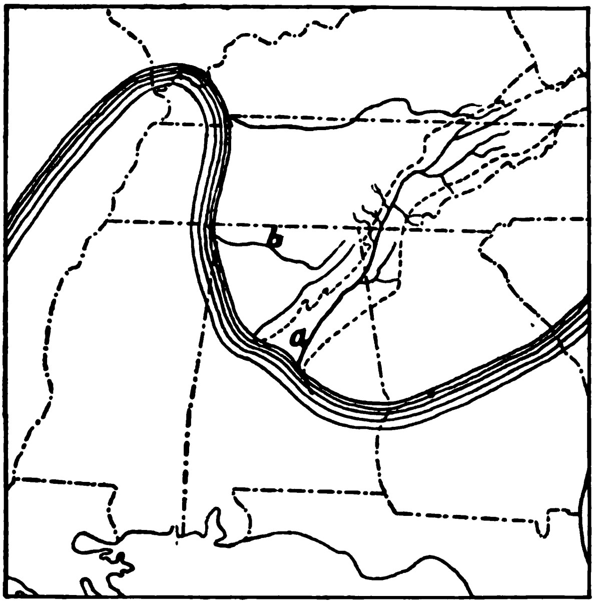

FIG. 165.—Shows the general position of the main drainage lines in the southern Appalachians at the close of the Cretaceous cycle of erosion. The lower part of stream *b* is made to follow the course of a portion of the present Tennessee.

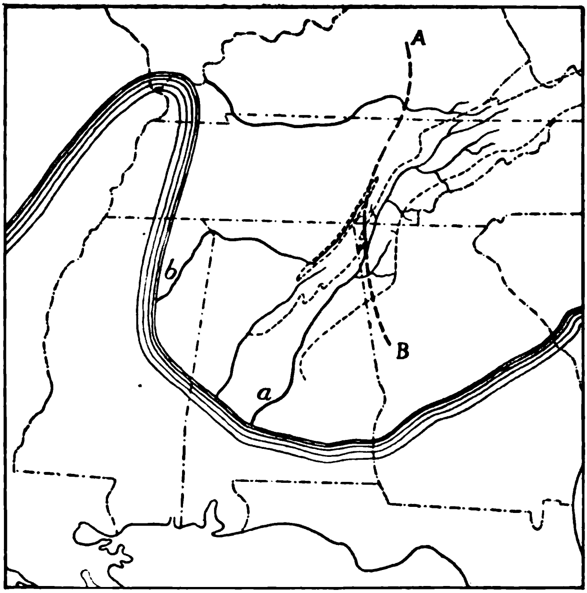

FIG. 166.—Shows the general position of the main drainage lines in the southern Appalachians, after the capture of the westerly tributaries of the Appalachian River by stream *b*. Compare Fig. 165.

of greatest elevation (*AB*, Fig. 166) was nearly parallel to the Appalachian River, and the effect of the differential uplift was to impose a greater task on this river (*a*, Fig. 166), which flowed along the axis of uplift, than upon the rivers which flowed westward and southwestward to the Mississippi embayment. The result was that the strongest of the

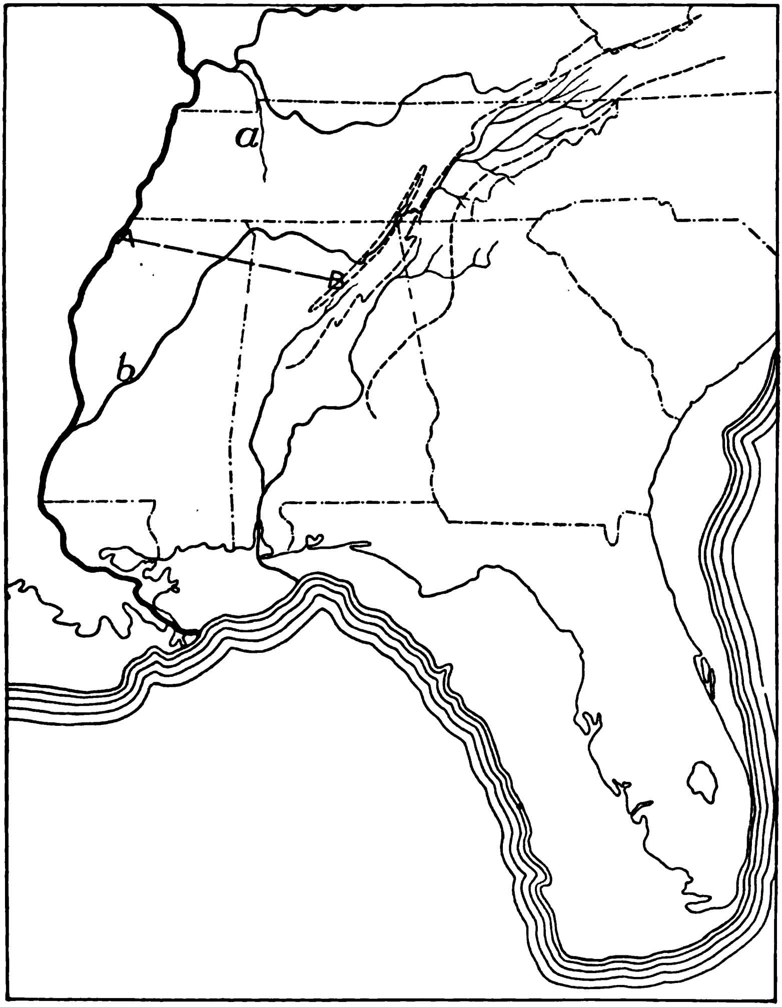

Fig. 167.—A stage later than that shown in Fig. 166. The sea is represented as having withdrawn from a considerable area which was submerged at earlier stages (Figs. 165, 166).

southwesterly flowing streams worked its head back into the drainage basin of the Appalachian River, and captured, one by one, the head-

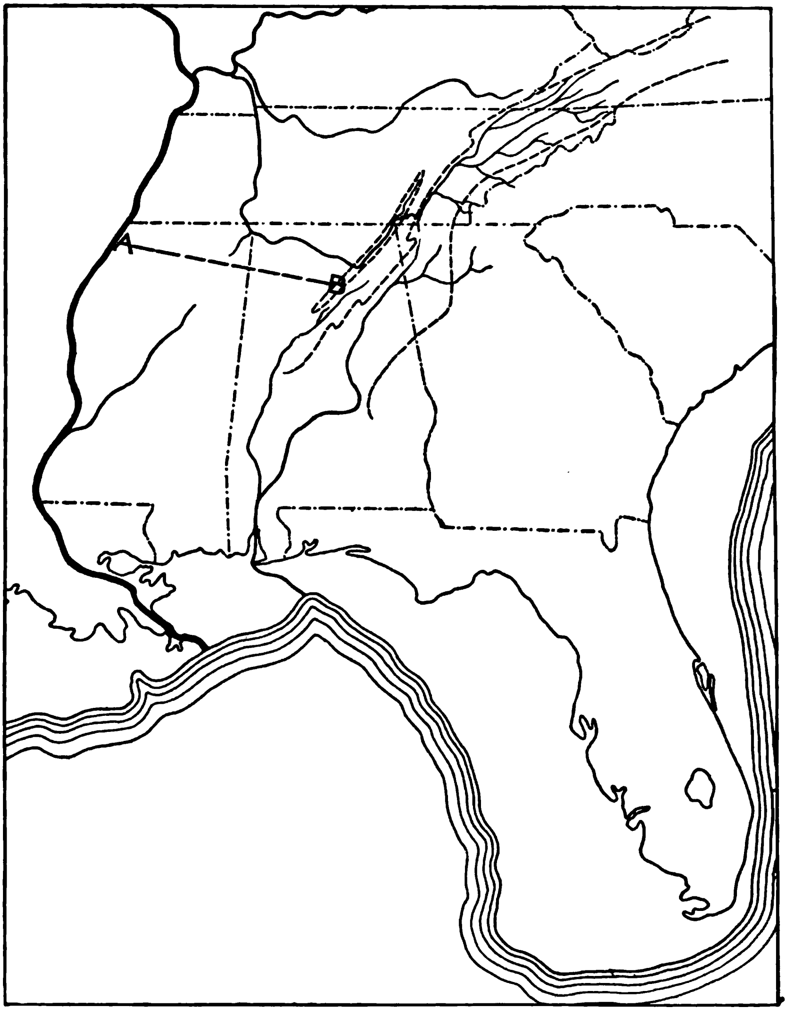

Fig. 168.—Shows the final change which resulted in the present course of the Tennessee. The land is represented as somewhat higher than now.[1]

waters of its westerly tributaries, establishing some such drainage relations as are shown in Fig. 166. Still later, after the land area of the region had been considerably extended by the withdrawal of the sea, the Appalachian River itself was reached by the invading stream, and its waters carried away to the Mississippi Bay by a course the lower part of which is thought to have corresponded approximately with the course of the present Black River (*b*, Fig. 167).

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[1] Figs. 165–168 are based on reports of Hayes, and Hayes and Campbell, already referred to. Drawn by E. S. Bastin.

Still later there was further deformation which caused additional changes in the drainage. The whole region was uplifted, relatively if not absolutely, but the uplift was differential, being greatest along the axis represented by *AB*, Fig. 167. The effect of the deformation was to stimulate the tributaries of the Ohio flowing north from this axis. Their growth was further accelerated by the weakness of the strata over which they ran. At the same time, the uplift to the south led the southwesterly flowing stream (*b*, Fig. 167) to discover relatively hard beds of rock in its lower course, and these beds retarded its down-cutting. The result was that a tributary of the Ohio (*a*, Fig. 167) finally tapped the main stream flowing to the southwest (*b*, Fig. 167) and carried its upper part over to the Ohio (Fig. 168). This was the beginning of the present Tennessee.

THE AGGRADATIONAL WORK OF RUNNING WATER.

Principles involved.—Since deposition results from the failure of transportation, the factors which control transportation also influence deposition. Transportation by streams is determined largely by velocity, and the most important factors influencing velocity are slope, volume, and load (p. 115). Of these the first two are usually of greater importance than the third.

A stream is said to be loaded when it has all the sediment it can carry; it is loaded with fine material when it has all the fine material it can carry, and with coarse material when it has all the coarse it can transport. A stream loaded with coarse material flows more swiftly than one loaded with fine, for a larger percentage of a stream's energy can be utilized in carrying fine material than coarse, and hence a larger percentage of the energy of a stream which carries a load of the latter will express itself in velocity.

Deposition takes place whenever a stream finds itself with more load than it can carry, and is an expression of the stream's refusal to remain overloaded. A stream may become overloaded in various ways. It might at first seem unnecessary to inquire whether a stream may be overloaded at its source, but the question is not necessarily to be answered in the negative. The source of a stream is not always a definite point. In a general way it may be said that the source of the normal stream is at that point in its valley where the bottom is as low as the ground-water level of the region. But since the ground-water level is not constant (p. 71) the source of a stream is likely to be farther up its valley in a wet season than in a dry one (p. 72). After a heavy shower, the run-off descends to the axis of the valley from the slopes on all sides, and temporarily the stream begins above the point which marks even its wet-season source. If under such circumstances the slopes about the head of the valley are notably steeper than the slope of the valley itself, as they frequently are, the water flowing down them may gather an amount of material which it cannot carry after it reaches the bottom of the valley. This may be the case at, or even above, the point which marks the source of the permanent stream. It is, therefore, possible for a stream to be overloaded at its source, if we take the source to be the point whence the water permanently flows. Deposition may, therefore, be taking place in a valley at the head of its permanent stream, or temporarily even in the valley above it.

Streams issuing from glaciers sometimes have more load than they can carry after they escape from the ice. If the stream be regarded as beginning at the point where it issues from beneath the ice it may be overloaded at its source.[1]

Under certain circumstances, a stream may overload itself. Thus if a stream loaded with coarse detritus reaches a portion of its valley where fine material is accessible in abundance, some of the velocity which is helping to carry the coarse may be used in picking up and carrying the fine. This reduces the velocity, and since the stream already had all the coarse material it could carry, reduction of velocity must result in deposition. It follows that when a stream fully loaded with coarse material picks up fine, it becomes overloaded, *so far as the coarse material is concerned.*

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[1] A question might be raised in this case as to what should be called the source. A spring issues from beneath the surface and flows away in a stream. The stream is said to begin where the water appears at the surface, though in some cases the water of the spring was a sub-surface stream before it reached the surface. Water escaping from beneath a glacier as a stream may likewise be considered a spring at the point of its issue.

Again, tributaries may overload their mains. While tributaries are usually smaller than their mains, they frequently have higher gradients, and the smaller stream of higher gradient may bring to the larger stream of lower gradient more material than the latter can carry away. Thus deposition may take place at the point of junction of tributaries with their mains. This may go so far as to pond the latter enough to cause its expansion into a river-lake. Lake Pepin, in the Mississippi River at the mouth of the Chippewa (in Wis.), is an example.

Streams may become overloaded by losing velocity or volume, or both. Decrease in velocity is brought about either by decrease in declivity or in volume. In general, streams have lower gradients and greater volumes in their lower courses than in their upper, and these two elements affect velocity in different ways. If the increase in volume be not enough to counterbalance the decrease in declivity, as is often the case, a stream which is loaded in its upper course will deposit in its lower. The decrease of velocity at the debouchure of a stream almost always leads to deposition.

Decrease in velocity as the result of decrease in volume is less common. When decrease in volume occurs, it may be the result of (1) evaporation, (2) the absorption of water into the bed of the stream, or (3) branching—the giving off of *distributaries.* While evaporation is going on everywhere, the diminution of a stream by this means is usually more than balanced by the increase from tributaries, rainfall, and springs; but in arid regions a very different condition of things sometimes exists. If mountains in an arid region be capped with snow, its melting supplies the streams during the melting season. As the streams flow out from the mountains through dry regions, they receive little or no increment from rainfall, tributaries, or springs, and evaporation reduces the volume of water, or even dissipates it altogether. Absorption of water into the bed of the stream often accompanies evaporation. Reduction of volume by evaporation and by absorption is especially common in arid regions. Wherever loaded streams are reduced in volume, whether by evaporation or absorption, deposition takes place.

The third way by which velocity is decreased as the result of decreasing volume is illustrated at the debouchures of many streams. Near the Gulf, for example, the Mississippi branches repeatedly (see Fig. 190). The same phenomena are often seen where one stream joins another (Fig. 169). Individually the distributaries are much smaller than the main stream before they separated from it, and because they are smaller their combined surfaces are greater, and the amount of energy consumed in the friction of flow is increased. The velocity of the water and its carrying power are, therefore, reduced. Thus the branching of streams gives rise to deposition, and where deposition takes place the gradient of the stream is reduced, and this occasions still further deposition. The sediment which fills up the channel and checks the flow finally compels the stream, or some part of it, to transgress its banks. Deposition, therefore, favors the development of distributaries, and the development of distributaries in turn favors deposition.

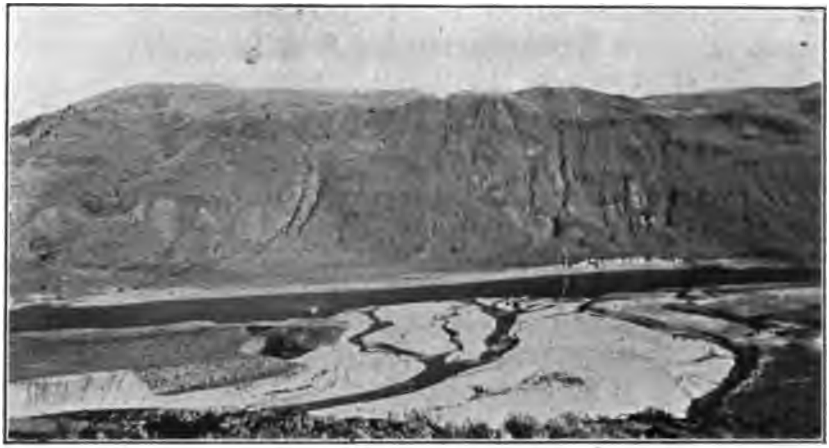

Fig. 169.—Delta of the Chelan River at its junction with the Columbia. Shows the tendency of streams to distribute where active deposition is in progress. (Willis, U. S. Geol. Surv.)

The foregoing statements make it clear that a stream may be eroding in one part of its valley while it is depositing in another, and that erosion may alternate with deposition in the same place, on account of fluctuations in volume, and, therefore, in velocity of the stream. It will be seen in the sequel that erosion and deposition may be taking place at the same time in the same part of the valley. The activities of a river are so nicely balanced that slight disturbance at one point causes disturbance at all points below.

The deposits.

Types.—Turning from the principles which underlie river deposition to the deposits themselves, they are found to occur in various situations. Running water usually descends from steeper slopes above to gentler slopes below, and ends at the sea, or in a lake or inland basin. Wherever there is a sudden decrease in its gradient, as at the base of a hill, ridge, or mountain, running water is likely to leave a large part of its load, building an *alluvial fan* or *cone* (Figs. 67, 68, and Pl. VI). Even where there is no sudden decrease in the gradient of a stream, there is likely to be a gradual one, and in spite of the fact that the increased volume of a stream in its lower course tends to overcome the effect of diminished gradient on velocity, deposition is likely to take place as the gradient is reduced. Deposits occasioned by the gradual reduction of a stream's velocity often have great extent in the direction of a stream's flow. They cover the flood plains of streams, making them *alluvial plains* (Fig. 73). When a stream reaches the sea or a lake its current is destroyed and its load dropped, unless taken in charge by the waves and currents of the standing water. Sediment accumulated in quantity at the debouchures of streams gives rise to *deltas* (Figs. 169, 187). Alluvial cones and fans, alluvial plains, and deltas, are the principal types of river deposits. Apart from these well-defined types there are *bars* in the channels of depositing streams, and much ill-defined alluvium which does not allow of ready classification.

Alluvial fans and cones.—The only distinction between the alluvial fan and the alluvial cone is one of slope, the cones (they are but half-cones at best) being steeper than the fans. Alluvial fans and cones have their most striking development where temporary torrents, occasioned by showers or the rapid melting of snow, issue from mountain ravines. Such streams usually carry heavy loads of detritus, the coarser part of which is likely to be deposited at the base of the mountain slope. Cones and fans built by such streams have a periodic rather than a steady growth.

At the beginning of its development the material of the alluvial cone is deposited much as in a talus cone (compare Fig. 170 with Figs. 67 and 68). Its deposition chokes the channel of the stream, and some of the water then seeks new courses to right and left of the apex of the deposit. This expands the area of deposition to right and left, while the water which flows over it lengthens it in the direction of flow.

The course and behavior of the water after reaching an alluvial cone is instructive. As its velocity is checked, deposition often takes place in the channel, diminishing its capacity. As the channel is filled up, the water tends to overflow on either side. The overflowing water, being shallow, has so little velocity that much of its load is dropped on either margin of the channel, building up levees. The water ever and anon breaks through the levees, giving rise to distributary streams, each of which aggrades its channel and builds its own miniature levees (Fig. 171). Not rarely this process of channel-filling and levee-building goes on until the channels of the little rivulets are above the general level of the cone on which they rest. The rivulet then runs in a groove on the crest of a little ridge. The channels on the surfaces of fans and cones are fewest and deepest at their heads, and more numerous and shallower below. In some cases the surface-water disappears altogether before the outer border of the fan is reached, by sinking into the débris.

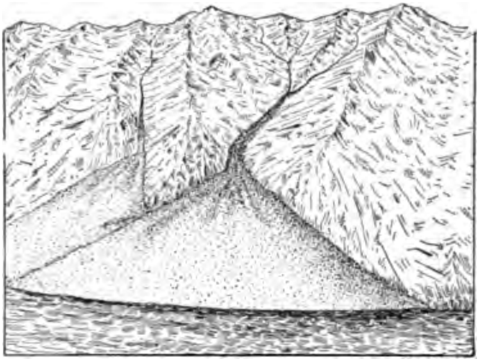

Fig. 170.—A talus cone. North Greenland Coast. The talus cone reaches the sea-level. Drawn from photograph.

Alluvial fans and cones have various forms, and often attain considerable dimensions. Their angles of slope depend on the amount of reduction of velocity which the depositing water suffers, and the amount and kind of load which it carries. The maximum slope of the cone is the angle at which the loose material involved will lie. The minimum slope of the fan, on the other hand, approaches horizontality. If many alluvial fans develop in promixity to one another, as at the base of a mountain range, they may expand laterally until they merge. A long succession of them may thus give rise to an extensive *alluvial piedmont plain,* or a *compound alluvial fan.* The lower edge of such a fan is often somewhat lobate. Such plains exist along the bases of many mountain ranges (Pl. VI), and may be seen in miniature even along low ridges.

A permanent stream, as well as a temporary one, may develop an alluvial fan at the base of a mountain slope; but since the mountain course of the former is likely to be less steep than that of the latter, its waters suffer a correspondingly less reduction of velocity at any one point. The fan of the permanent stream is therefore likely to be relatively flat, and to stretch far down the valley. Such fans grade into valley plains. From the general principles already discussed, it is clear that well-developed fans go with relatively youthful stages of erosion, and belong normally to the upper parts of drainage lines.

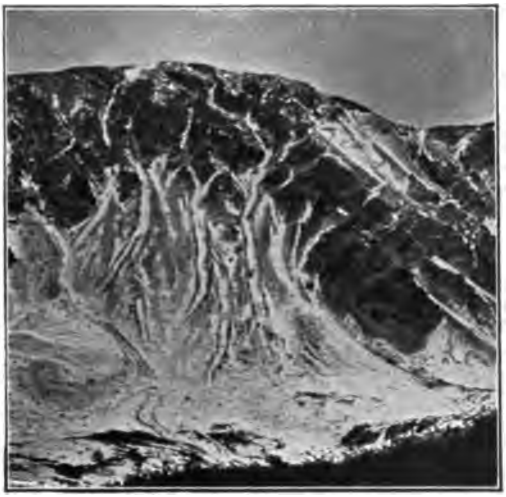

Fig. 171.—Miniature levees on an alluvial cone. Slope of Gray Peak, Colo. (R. T. Chamberlin.)

Ill-defined alluvium.—There is a widespread mantle of alluvial material deposited by running water which was not organized into distinct streams. The water which runs down smooth slopes in sheets during showers carries fine earthy matter, as well as some that is coarser. These materials are largely deposited at the bases of the slopes, forming basal accumulations of greater or less extent, comparable in origin to alluvial fans. A relatively small amount of the slope wash is carried far out from the base of the declivities. It is not easy to realize the extent to which this process is taking place. There is hardly a slope without loose material, and there is hardly an acre of low land below a slope on which running water has not deposited sediment washed down from above. When it is remembered that this is as true of gentle slopes and their surroundings as of steep slopes, though perhaps not to the same extent, and that a very large part of the earth's surface is made up of sensible slopes, or of flats at their bases, some idea of the aggregate effect may be gained.

There is another way of looking at the same question. Earthy matter is being continually transferred from land to sea, and chiefly from high land. Rarely does it start from any point distant from the shore and move uninterruptedly to it. It is transported a short distance and lodged, to be again picked up, carried forward another step in its journey, and lodged again. For a very large part of the earth's surface it would be true to say that its mantle rock is material in transit from higher land to the sea.

Alluvial plains.—Most streams, whether heading in mountains or not, have gentler gradients in their lower courses than in their upper, and in spite of increasing volume are usually unable to carry to their debouchures all the material gathered above. The excess of load is dropped chiefly on the flood-plains of the streams and constitutes them *alluvial plains*.

The making of an alluvial plain usually involves both erosion and deposition. When a stream has cut its channel to grade, downward erosion ceases, or more exactly, downward cutting is, on the average, counterbalanced by deposition. So long as a stream is cutting downward rapidly, it carries away whatever débris descends the side slopes. When it approaches grade, the débris which descends the side slopes tends to accumulate at their bases, and the V-shaped cross-section of the valley becomes U-shaped (see Fig. 172). At about the same time the stream begins to meander, for, having lost something of its former velocity, it is more easily turned from side to side. As it begins to meander, it widens the bottom of its valley. This is the initial stage in the development of the valley flat (2 and 3, Fig. 172). In its meandering the stream encroaches on the talus accumulations at the bases of its valley's slopes. The side-cutting may remove all the loose débris and even undercut the bluff as at *a*, Fig. 173. The stream's meanders shift their positions from time to time so that the valley flat is successively widened at different points. By lateral planation, therefore, a stream tends to develop a flat as soon as it reaches grade. This is the initial part of erosion in the making of a river flat, but a flat developed by erosion alone is not an alluvial plain.

So soon as the flat developed by a stream exceeds the width of its channel, the water (except in times of flood) does not cover it all at the same time. On any part which it temporarily abandons, some débris (alluvium) is likely to be left. This deposit of alluvium constitutes the valley flat an alluvial plain (Fig. 174). It will be seen that the valley flat is commonly an alluvial plain from the beginning.

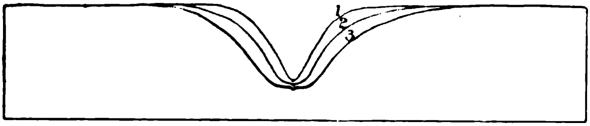

Fig. 172.—Diagram illustrating the transformation of a V-shaped valley into a U-shaped valley.

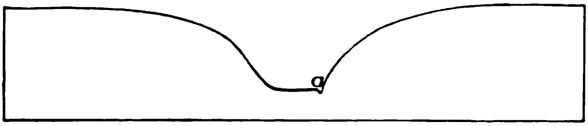

Fig. 173.—Diagram to illustrate the widening of a valley flat by erosion. Compare 3, Fig. 172.

Once the valley flat and alluvial plain are begun, their further development is easily followed. The stream in flood overflows the banks of its channel. The velocity of the overflowing water is reduced, and if it has much load a part of it will be dropped and the plain aggraded. Meantime meandering and lateral planation continue. Thus the flood-plain is widened by erosion, and aggraded by alluviation, the two processes going on simultaneously.

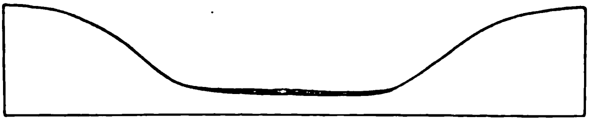

Fig. 174.—An alluvial plain. The diagram suggests the relative importance of lateral planation and alluviation in the development of the flat.

Flood-plains, chiefly the result of planation, but partly of aggradation, are a normal feature of river valleys, after a certain stage of development has been reached. This stage is that at which downward erosion becomes slight in comparison with lateral erosion. It follows that an alluvial plain normally begins its development where the valley is first brought to grade, that is, in its lower course. As the development of the valley goes on, the head of the flood-plain advances up-stream, and at the same time its older parts become wider.

Flood-plains due to alluviation only.—Exceptionally, an alluvial plain is developed by deposition only. Thus if a stream becomes overloaded while its valley is still narrow, as sometimes happens, deposition follows, and, as aggradation proceeds, the narrow valley acquires a progressively wider bottom (Fig. 175). Wide valley plains are sometimes developed in this way. Flood plains developed wholly by alluviation are sometimes formed under conditions which are independent of the stage of a valley's development. Thus if a stream suddenly acquires an exceptional supply of detritus in its upper course, the development of an alluvial plain begins immediately below the point of overloading.

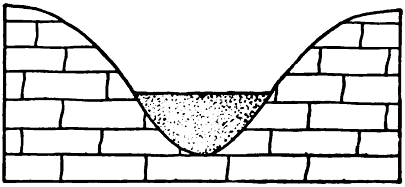

FIG. 175.—Diagrammatic representation of a flood plain developed by alluviation only.

The overload might be acquired in various ways. (1) If a stream taps another (piracy) which carries a large quantity of sediment, carrying off both water and sediment to a channel with a lower gradient, deposition may take place where, under the earlier conditions, there was none. (2) Again, when a stream cuts through a barrier near its head waters, its velocity, and, therefore, its eroding power, may be so increased in its upper course that sediment enough is acquired to occasion deposition below, where none took place before. (3) In working back through formations of varying degrees of resistance, a stream's head may presently reach a formation or a region which yields abundant sediment, even though there was no especial barrier below. (4) If an advancing glacier should reach the head waters of a stream, its discharge to the stream would greatly increase the load of the latter, and, although its volume would be augmented at the same time, deposition might result. As a matter of fact, streams carrying glacial drainage are usually aggrading streams. In general, anything which greatly increases the load of a stream near its head is likely to cause deposition, and so the development of a flood plain, at some point farther down the valley.

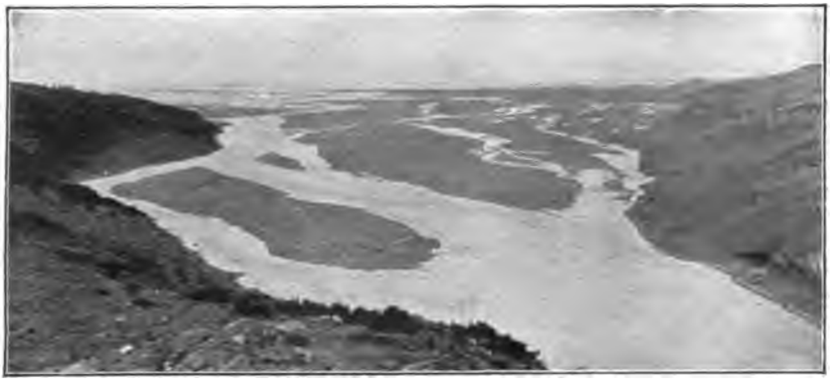

FIG. 176.—Anastomosing of a depositing stream. Yahtse River, Alaska. (Russell, U. S. Geol. Surv.)

Streams which are actively aggrading their valleys are likely to anastomose (Figs. 176, 177). This results from the filling of the channels until they are too small to accommodate all the water. The latter then breaks out of the channel at few or many points. The new channels thus established suffer the same fate.

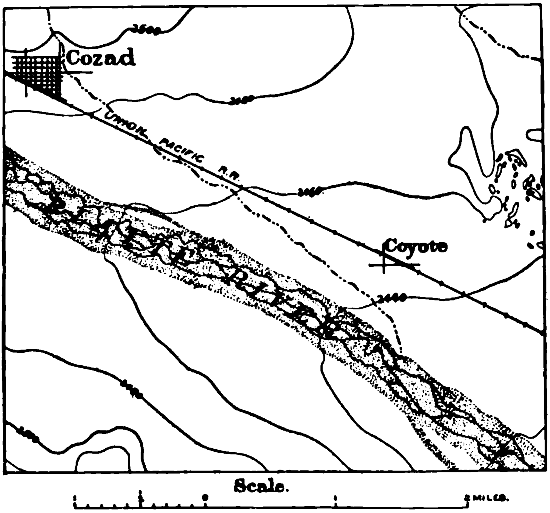

FIG. 177.—Anastomosing of the Platte Rive, Dawson Co., Neb. (U. S. Geol. Surv.)

Flood-plains due to obstructions.—Again, any obstacle in a stream's course is likely to cause deposition above. Thus dams built across rivers entail the deposition of sediment above. Where a stream flows over the outcropping edges of strata of different strength, the more resistant serve, in some sense, as dams. Above them the stream cuts its bed to a low gradient, and, becoming sluggish, drops more or less of the detritus brought down from above. Obstacles of any sort across a stream's channel, therefore, favor the development of alluvial plains.

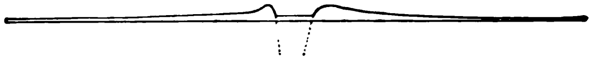

FIG. 178.—The levees of the Mississippi in cross-section, 4 miles north of Donaldsonville, La. Vertical scale × 50. The horizontal line in the diagram represents sea-level. The bottom of the channel at this point is far below sea-level.

Levees.—As the stream in flood escapes its channel and overspreads its plain, its immediate banks are the site of active deposition, for it is here that the velocity of the overflowing water is first notably checked. On the banks of the channel, therefore, low alluvial ridges, called *natural levees*, are built up (Fig. 178, and Pl. XV). They may be narrow, or hundreds of feet in width, and are often several feet above the plains behind them, giving the latter a slope away from the channel of the stream. They are sometimes high enough t control the courses of tributary streams, as shown by numerous tributaries to the Mississippi below the Ohio. The Yazoo, for example, flows some 200 miles on the flood-plain of the Mississippi before it joins that river near Vicksburg. The levees even become divides, directing drainage away from the streams they guard (Pl. XV). Streams sometimes build levees faster than their tributaries aggrade their channels. The latter are then ponded, giving rise to lakes. The lakes on the lower courses of the tributaries to the Red River of Louisiana are examples.[1] They are sometimes built up above their natural level and kept in repair by human agency so as to confine the streams in time of flood. This is a source of danger unless they be steadily maintained, for the breaking of such levees often occasions great destruction. A case in point is the breaking of the levees of the Mississippi near New Orleans in 1890. The water broke through the levees at the Nita and Martinez crevasses (Fig. 187) and flowed eastward (from the former) with a current of 15 miles per hour, spreading destruction in its

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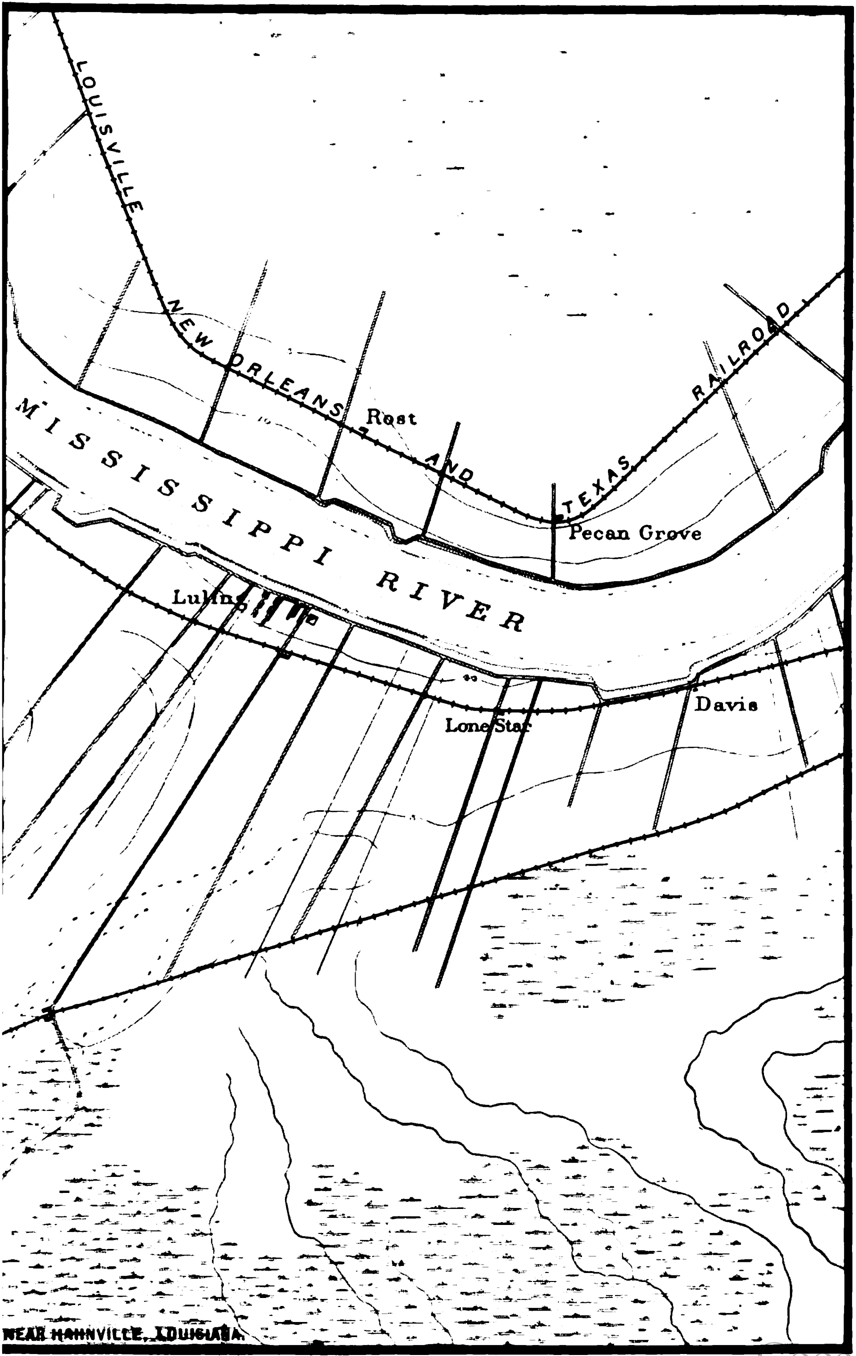
[1] Davis. Science, Vol. X, p. 142, 1887.


LOUISVILLE NEW ORLEANS AND TEXAS RAILROAD
MISSISSIPPI RIVER
Rost
Pecan Grove
Lulings
Lone Star
Davis
NEAR HAHNVILLE, LOUISIANA.

U. S. Geol. Surv.

Scale, 1+ mile per inch.

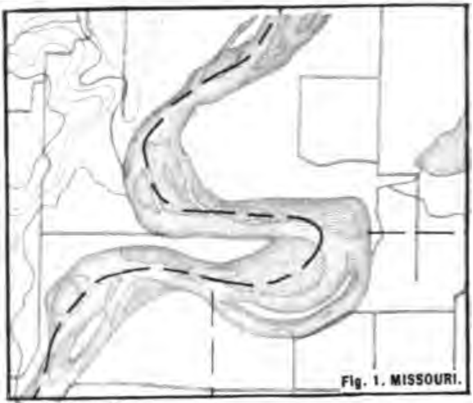


Fig. 1. MISSOURI.

U. S. Geol. Surv.

Scale, 2+ miles per inch.

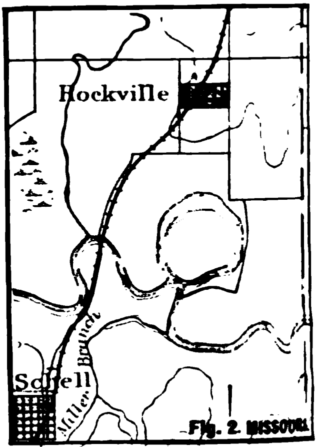


Fig. 2. MISSOURI.

U. S. Geol. Surv.

Scale, 2 miles per inch.

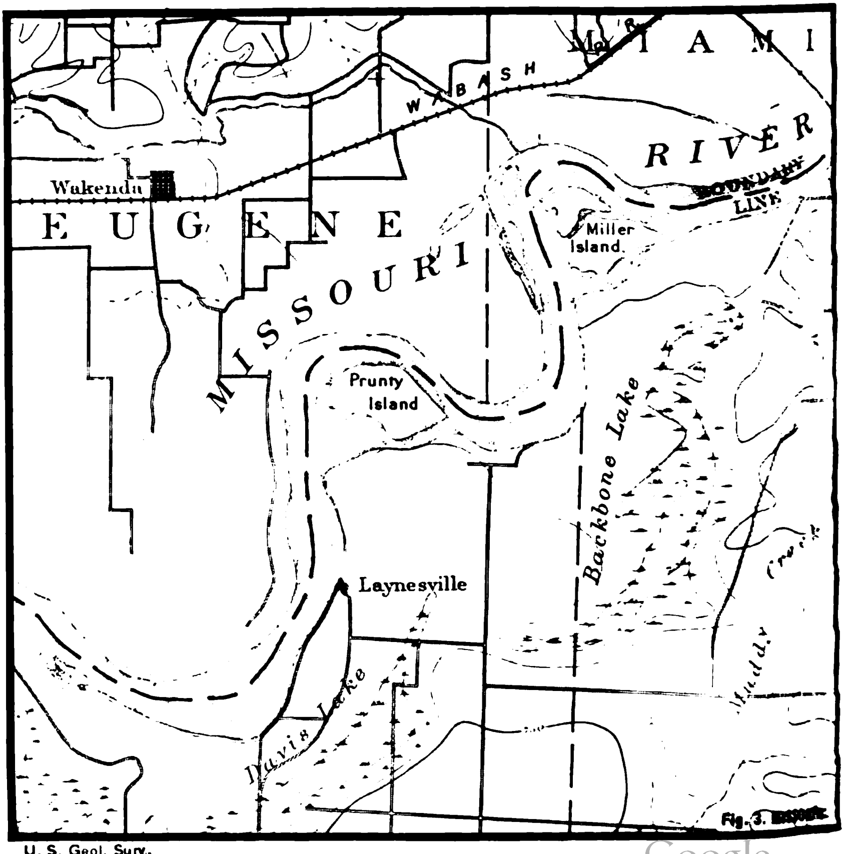


Fig. 3. MISSOURI.

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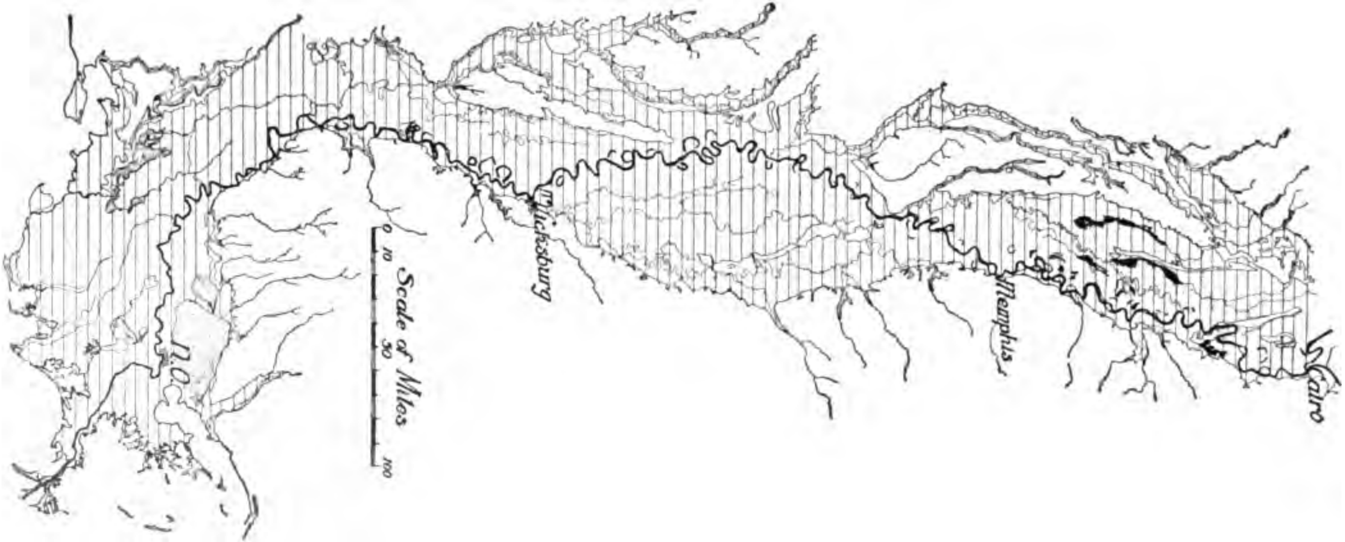

FIG. 179.—Flood-plain of the Mississippi River south of the mouth of the Ohio. (From charts of the Miss. Riv. Commission.) (Page 189.)

path. The water flowed eastward through Lakes Pontchartrain and Borgne, and entered Mobile Bay with such volume, velocity, and load of mud, as to destroy for a time the oyster and fish industries of that locality.[1]

Flood-plain meanders. Cut-and-fill.—A stream with an alluvial plain is likely to meander widely (Pl. XVI). In general terms this may be said to be the result of low velocity, which allows it to be easily turned aside. Were the course of such a stream made straight, it would soon become crooked again. The manner of change is illustrated by Figs. 180 and 181. If the banks be less resistant at some points than at others, as is always the case, the stream will cut in at those points. If the configuration of the channel is such as to direct a current against a given point, *a* (Fig. 180), the result is the same, even without inequality of material. Once a curve in the bank is started, it is increased by the current which is directed into it. Furthermore, as the current issues from the curve, it impinges against the opposite bank and develops a curve at that point. The water issuing from this curve develops another, and so on.

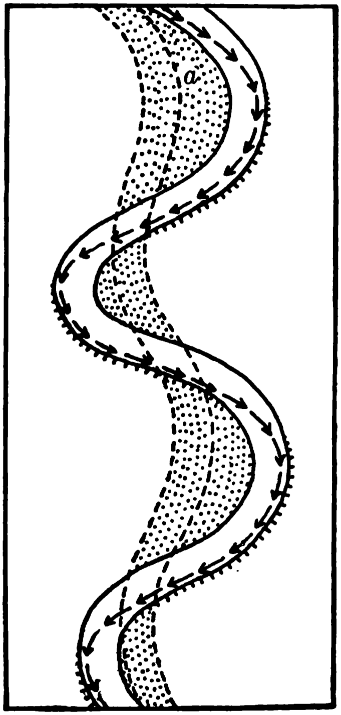

FIG. 180.—Diagram illustrating an early stage in the development of meanders. The shaded part represents the area over which the stream has worked.

Once started, the curves or meanders tend to become more and more pronounced (compare Figs. 180 and 181). In the case represented by Fig. 1, Plate XVI (Missouri River near Brunswick, Mo.) the narrow neck of land between curves is almost cut through. When this is accomplished, the stream will abandon its wide curve. A later stage in the process is shown in Fig. 2, Plate XVI (the Osage River near Schell, Mo.).

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[1] L. C. Johnson. Bull. Geol. Soc. Am., Vol. II, pp. 20–25, 1891.

The straightening of the channel is often accomplished in another way. Even before the meanders reach the stage represented by Fig. 1, Plate XVI, the position of the channel becomes unstable. In time of flood, the whole flat is covered with flowing water. The greater depth of water in the channel tends to give it a velocity greater than that of the water on the flat outside. But the distance from *a* to *c via b* (Fig. 181) is much greater than that in a direct line. It follows that the slope from *a* to *c* direct is greater than that by way of *b*. If the current between *a* and *c* in time of flood be strong enough to erode, it may deepen its bed, and thereby increase the volume of water following this course. The increased volume gives increased velocity, and the result may be the opening of a channel between *a* and *c* direct. The channel may be worn so deep that when the flood subsides, the stream will follow it. So long as the abandoned channel-curve remains unfilled with sediment, it is often called a *cut-off*. If it contains standing

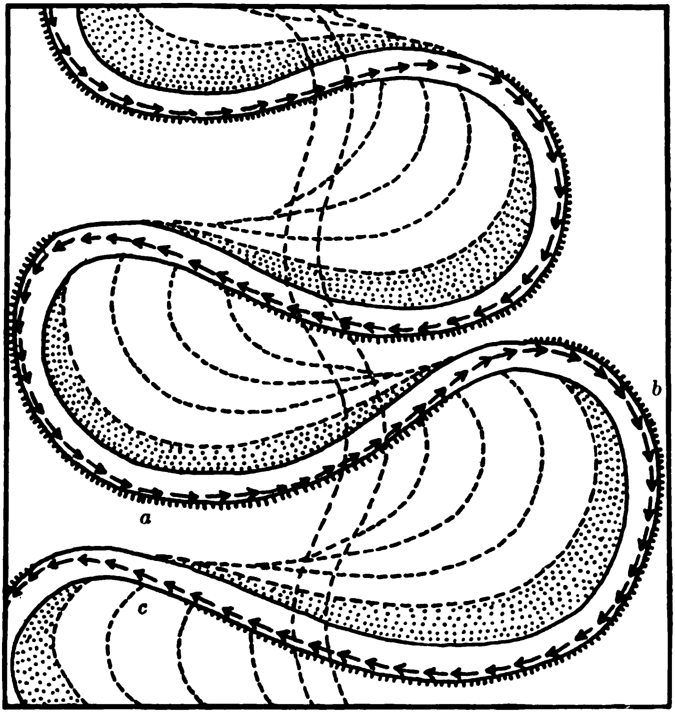

FIG. 181.—Diagram illustrating later stages in the development of meanders.

water and has the proper form, it is called an *ox-bow lake* (Fig. 182), or sometimes a *bayou*. The water-filled portions are not always bows

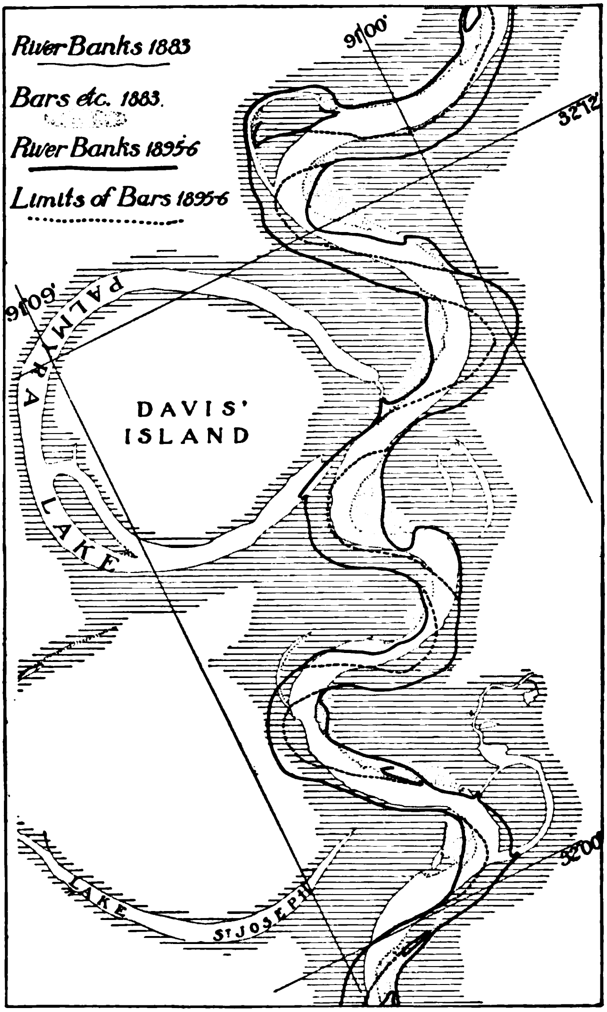

FIG. 182.—Meanders and cut-offs (ox-bow lakes) in the Mississippi Valley a little below Vicksburg. The figure also shows the migration of meanders down-stream, and their tendency to increase themselves. (From charts Nos. 18 and 19, Mississippi River Commission.)

(Fig. 183, Osage River, near Butler, Mo.). Cut-offs, with or without standing water, are of common occurrence along most rivers with wide plains. Meandering is not confined to streams which are near sea-level. Even small creeks at high altitudes may meander, if so situated as to have slight velocity. Trout Creek in the Yellowstone Park (Fig. 184) is an example.

There seems to be some relation between the width of the belt within which a stream meanders, and the width of the stream itself. Recently it has been estimated that the ratio between them is 18 : 1.[1]

During the development of the meanders it is to be noted that lateral planation on the one side of a stream is accompanied by deposition on the other. This is *cut-and-fill.* The sediment eroded from the curve which is concave toward the stream is shifted down-stream, while that deposited in the curve which is convex toward the stream is brought down from above. Thus even in the development of meanders, the material which is dislodged is shifted down-stream. Since the current directed against the down-stream side of a growing meander is on the average stronger than that directed against the opposite side, the meander itself has a tendency to migrate down-stream (Fig. 182).

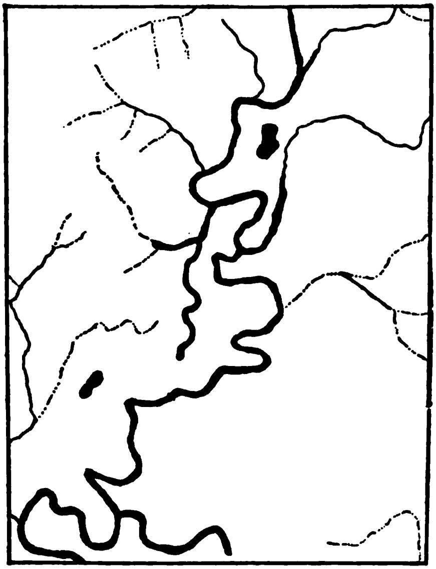

FIG. 183.—Bayou Lakes, Osage River, near Butler, Mo.

In their evolution, the curves of a stream's channel often reach and undermine the valley bluff (Pl. VII). Since the meanders are, on the average, shifted down-stream individually, and since meanders are frequently developed in new places, it follows that a meandering stream tends to widen its valley throughout. Widening is also effected in other ways, for a stream with a flood-plain sometimes abandons its channel altogether for miles at a stretch, and the new course chosen may be against one of the bluffs of the valley. Such changes are most likely to take place where deposition along channel and levees has brought the part of the flood-plain (though not necessarily the bottom of the channel) adjacent to the stream above the level of that farther from it (Fig. 178). The change is likely to be effected in time of flood.

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[1] Jefferson. Nat'l Geog. Mag., Vol. XIII, pp. 373–84.

Flood-plains often attain great size. That of the Mississippi below the Ohio (Fig. 179) has a width ranging from rather more than 20 miles at Helena (Ark.), to something like 80 miles in the latitude of Greenville (Miss.).[1] Below the Ohio its area is something like 30,000 square miles, and its entire area has been estimated at about 50,000 square miles.[2]

Theoretically, the rotation of the earth should affect the erosion of streams, increasing it on the one bank (the right in the northern hemisphere and the left in the southern) and decreasing it on the other. The streams doubtless accommodate themselves to the rotation of the earth in the original development of their gradation-plains and flood-plains, and the later effects of rotation are usually inconspicuous

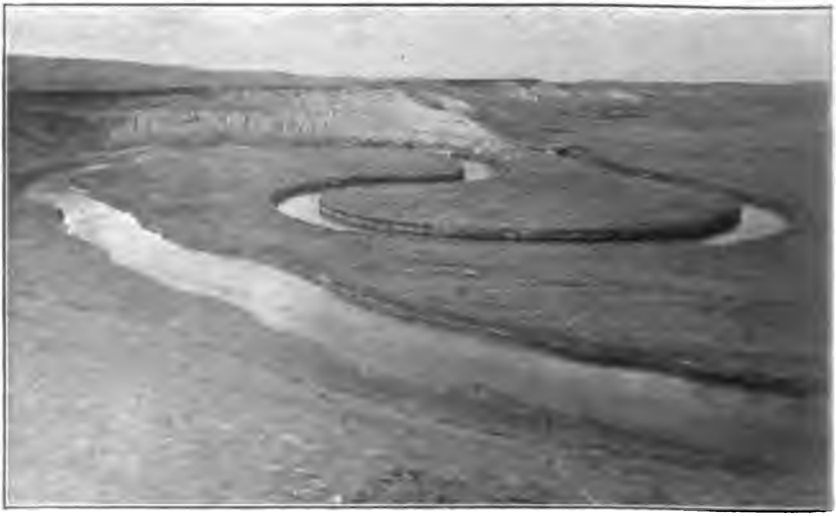

FIG. 184.—Meanders of Trout Creek, Yellowstone Park. (Walcott, U. S. Geol. Sur

Scour-and-fill.—It has already been shown that aggrading str cut laterally at the same time that they build up their plains. It now to be added that they periodically deepen their channels to a notable extent, and that the deepening of the channel takes place at the very time when the flood-plain is being aggraded. In other words, the stream in flood aggrades its plain, and degrades its channel. This follows from the fact that the current is sluggish in the former position, where the water is shallow, and rapid in the latter, where it is deep. When the flood subsides, the channel, deepened while the current was torrential, is filled again by the feebler current which follows. This alternate deepening and filling is known as scour-and-fill. It is well illustrated by the Missouri River. At Nebraska City, scour is believed to occasionally reach depths of 70 to 90 feet.[1] At Blair, about 25 miles above Omaha, the same river is believed to cut to bed-rock (about 40 feet below the bottom of the channel in low water) twice a year, that is, during floods.[2] Fig. 185 shows the changes recorded in the channel of the river at this point during the year 1883. It shows that the scour-and-fill during this year amounted to almost 40 feet. All

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[1] According to map published by the Mississippi River Commission in 1887.
[2] Russell. Rivers of North America, p. 114.
[3] Gilbert. Am. Jour. Sci., Vol. XXVII, 1884, pp. 427–34.

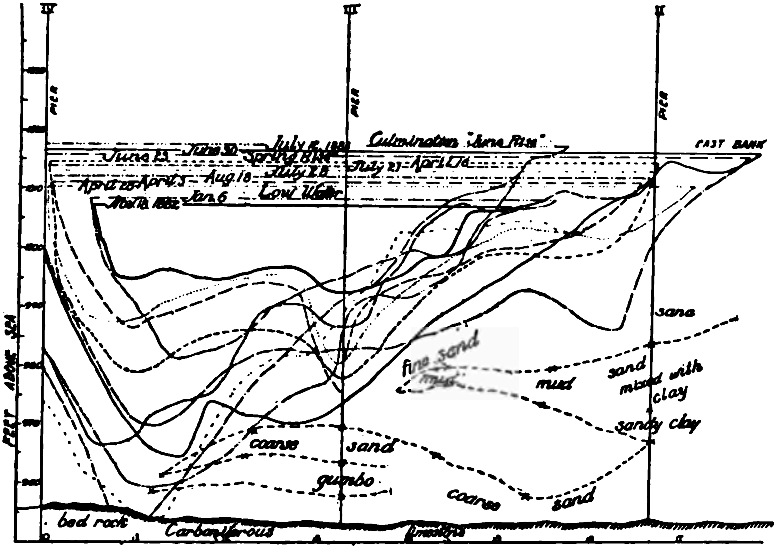

Fig. 185.—Diagram illustrating scour and fill in the Missouri River. A record of soundings at Blair Bridge (near Omaha), 1883. Shows also the cross-sections of the river at various rates. (Todd, Bull. 158, U. S. Geol. Surv.)

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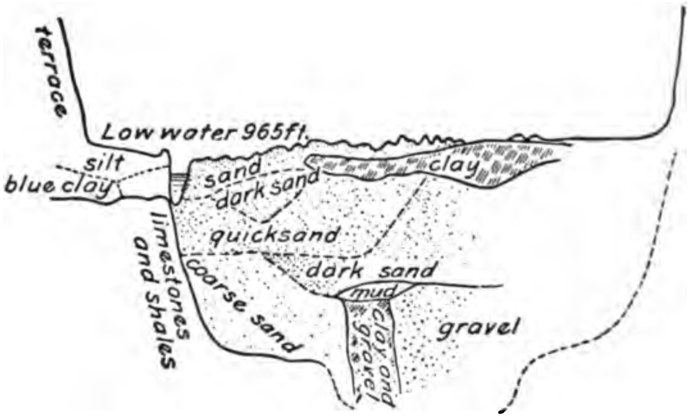
[1] Cooley. Rept. U. S. Engineers for 1879–80, Pt. II, pp. 1060 and 1071.

[2] Gerber. Cited by Todd. Bull. 158, U. S. Geol. Surv., pp. 150, 151.

streams similarly situated do a like work. The material thus eroded is shifted down-stream, some of it for short distances only, and some of it to the sea. Even an aggrading stream therefore is not without erosive activity; it is a stream whose fill exceeds its scour, not one which has ceased to erode.

Materials of the flood-plain.—As a result of its varying velocities in flood and low water, a stream may deposit coarse material at one time and fine at another. A similar sequence of deposits takes place in the flood-plain of a meandering stream, irrespective of floods. Flood-plain deposits are often therefore very heterogeneous, as shown in Fig. 186, which represents the constitution of the alluvium of the Missouri River at Omaha. The deposits of the streams range from the finest clay, through sand to gravel, and even bowlders. In general they become finer down-stream. In a given plain, they are usually coarser below and finer above.



Omaha Section

FIG. 186.—Diagram to show the heterogeneous character of alluvial deposits. (Todd, Bull. 158, U. S. Geol. Surv.)

Topography of the flood-plain.—The flood-plain is nearly, but not altogether, flat. It has a gentle slope down-stream, and often for a distance from the sides toward the center (Fig. 174). This latter slope is the result of deposition by waters descending to the plain from the sides. It is destroyed wherever a meandering stream reaches its bluffs. When levees are well developed, there is a slope from them toward the sides of the valley (Fig. 178), but it rarely continues to the limiting bluffs. Since a stream with a well-developed flat frequently shifts its course, old levees and abandoned channels lend variety to the topography of the flood-plain.

The topographic adjustment of tributaries.[1] — The meandering and shifting of a main stream affects its tributaries. If a main stream swings against the bluff through which a tributary enters, the latter brings its channel into topographic adjustment by lowering its end to the level of the main. If now the main stream opposite the tributary swings to the other side of its valley, the tributary must make

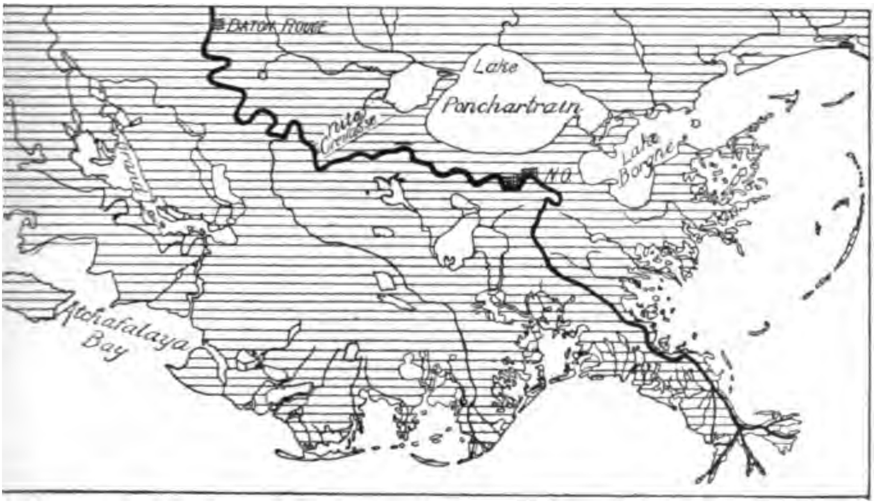


Fig. 187.—A general view of the Mississippi delta.

ts way across the flat with a very low gradient. Not only this, but he flat of the main valley through which the tributary must flow is ikely to be aggraded by the main in time of flood. The result is that he tributary stream becomes an aggrading stream at its debouchure, nd topographic adjustment is not established until it has filled up he lower end of its valley to some notable extent. The filling of the ower end of the tributary likewise affects the lower ends of its lower ributaries.

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[1] Chamberlin. Jour. of Geol., Vol. X, pp. 747–754.

If the main stream again swings over to the point where the tributary issues from its valley, the tributary stream and all its affected tributaries again become eroding streams. Thus scour-and-fill are not confined to the valley of the main stream.

River-lakes. — While rivers are in general hostile to lakes, they sometimes give origin to them. Oxbow lakes (Fig. 182 and Pl. XVI), due to the cut-offs of meandering streams, have already been referred to. Lakes formed in the same way have other forms (Pl. XVI and Fig. 183). Rivers also give rise to lakes through the deposits they make. If a main stream obstructs its tributaries by deposition at their debouchures, their lower courses are ponded and converted into lakes. The lakes along the tributaries to the Red River of Louisiana have already been cited as examples. If a tributary brings more load to its main than the latter can carry away, the detritus constitutes a partial dam, ponding the river and causing it to expand into a lake above. Such is the origin of Lake Pepin already referred to. In mountain regions, the alluvial cones of tributary valleys sometimes pond their mains.

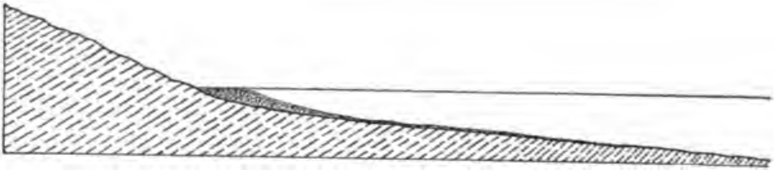

Fig. 188.—Longitudinal section of an incipient delta made of coarse material.

Rivers may be dammed in other ways, as by lava flows, by landslides, by glacial drift, etc. In all such cases, lakes may come into existence, but they are not due primarily to the activity of the river itself.

Deltas.[1]—Where a stream enters standing water, or a slower stream, a special form of plain, *the delta,* is sometimes built up (Figs. 169, 187, and 188). Deltas and alluvial fans have much in common, and their only notable differences are those imposed by the differences in the conditions of deposition. The current of the stream is checked, but not altogether stopped, at its immediate debouchure. If it carries abundant sediment, much of it will be promptly dropped where the decrease in velocity is first felt. Such flow as there is beyond the debouchure is not confined to a definite channel, and the deposits made are therefore spread more or less on either side of the line which represents the continuation of the stream's course.

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[1] For an excellent discussion of deltas, see Gilbert, Fifth Ann. Rept. U. S. Geol. Surv., pp. 104–8. Also Lake Bonneville, Monograph I, U. S. Geol. Surv. (same article).

As the depth of the water into which the stream flows increases, the current diminishes. Out to the point where the depth of the standing water is less than the depth of the current, the latter affects the bottom, and the surface of the deposits made slopes gently seaward; but where the depth of the standing water is such that the projected stream current is ineffective at the bottom, all the load rolled along the bottom is dropped, and a depositional slope is established (Fig. 188), its upper edge being below sea-level by an amount corresponding roughly to the depth of the current which brings the detritus. The outer slope is relatively steep and well-defined where the detritus is coarse, and relatively gentle and ill-defined where it is fine. Thus the stream tends to construct a sort of platform in the water just beyond its debouchure. The successive deposits on the outer abrupt slope will dip conformably with its surface (Fig. 189). The finest sediment will be carried beyond the steep slope, and conform to the topography of the bottom beyond (*c*, Fig. 189).

At the beginning, the top of the delta platform is at the level of the bottom of the stream's channel at the point of debouchure, but it is gradually aggraded as water continues to flow over it. Its landward margin is presently built up to sea-level and then above it, and as the delta grows the delta-land is extended seaward (compare Figs. 188 and 189). At the same time the channel of the stream above the original head of the delta is aggraded, for the current there is checked by the aggradation of the delta. Thus alluvial deposits continuous with the delta are extended landward.

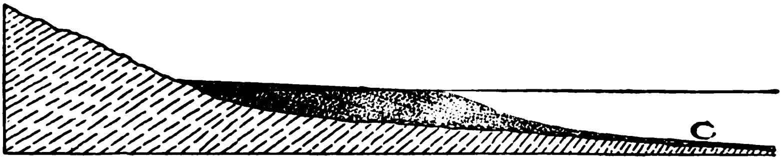


FIG. 189.—Longitudinal section of the delta at a later stage of development.

The projection of the direction of the lower end of the stream may be said to be the axis of the extra-debouchure current. From this axis, where the flow is strongest, the movement diverges more or less to right and left. Since the velocity of the diverging water is reduced more rapidly than that of the water which follows the axis of flow, deposition is likely to take place faster on either side of the axis than along the axis itself. The result is that the extra-debouchure current tends to build up levee-like ridges on either side, making a sort of sluice-way for itself. This sluice-way is gradually extended seaward, and at the same time gradually filled. As its capacity is reduced, more and more of the water flows over its sides. Presently the escape of the water over the little side-levees will develop a break at some point, and a line of distinct flow then diverges from the main current. This distributary repeats the history of its main. Thus the processes of levee-building, channel-filling, and levee-breaking follow one another,

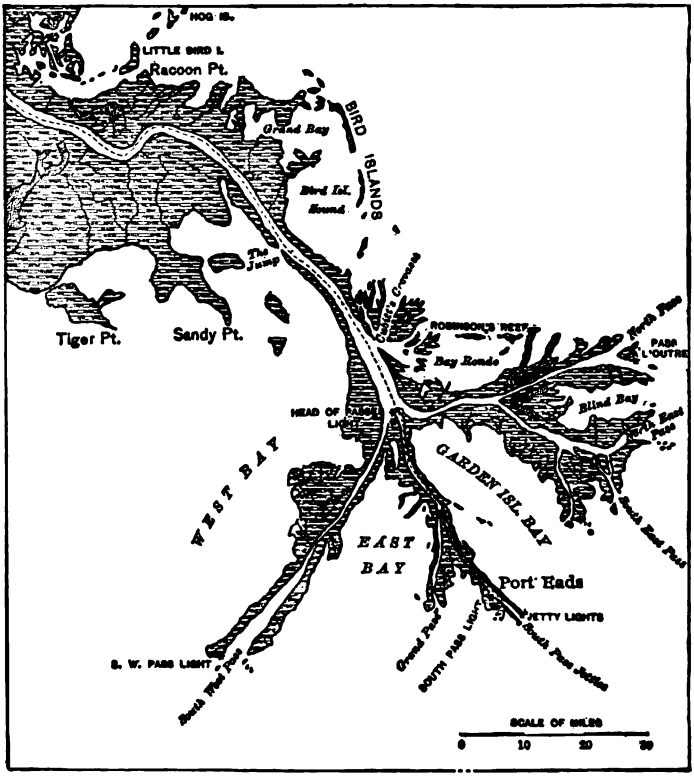
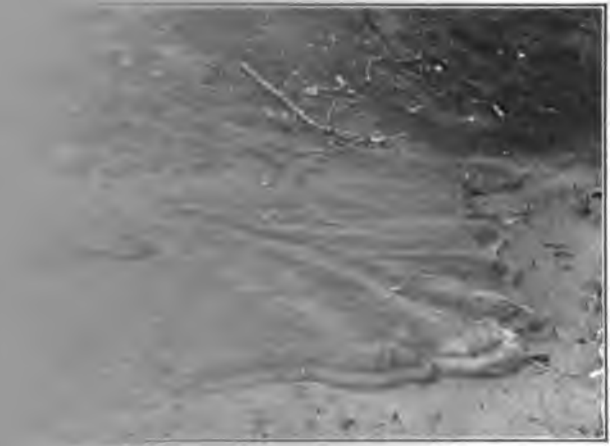


FIG. 190.—The terminus of the delta of the Mississippi. (C. and G. Survey.)

as shown in Fig. 190 is developed. is not simply in the line of extension more or less semicircular area, the center below the position of the debouchure begun. At any stage in its development or less crenate (Fig. 191), or character- the projections corresponding to the of the latest streams flowing across it. lobes (Fig. 190), and of groups of the of the shape of the Greek letter from but the resemblance in form between a Greek letter is not striking. Deltas are



191.—A miniature delta.

and in such cases their forms are predeter- The head of a delta is sometimes arbitrarily the first distributaries are given off. Since time, the definition can hardly be accepted. of the Mississippi delta is about 200 miles above much farther north.

shown in Fig. 189, shows its history. At the river discharges its sediment across that built. The sediment rolled at the bottom on reaching the steep slope, and constitutes the inclined *fore-set* beds shown in Fig. 189. The material in suspension is carried farther, settles more gradually, and constitutes the *bottom-set* beds (*c*, Fig. 189). In time the bottom-set beds, originally deposited some distance beyond the debouchure, may come to be overlain by the fore-set beds, deposited at a later time. While the fore-set beds are being deposited on the steep slopes of the delta, and the bottom-set beds beyond, deposition is also taking place on the top of the delta. These *top-set* beds are laid down in a nearly horizontal position, and their seaward margin is gradually extended. Thus the delta comes to have the threefold structure shown in Fig. 189.

That part of the delta which is above the abrupt slope of its front corresponds in all essentials to an alluvial fan; but the delta as a whole differs from the fan in its abrupt and crenate or digitate margin.

It is to be noted that the delta is not wholly the product of a stream's activity. The stream supplies the material, but the lake or sea renders at least passive assistance in its disposition. Not all rivers opening into the sea build deltas, and their failure is often the result of waves or shore currents which carry off the river sediment. Deltas are, however, sometimes formed in tidal seas, as at the debouchures of the Yukon; the Mackenzie, where the tidal range is three feet; the Niger, where the range is four feet; the Hoang-Ho, where the range is eight feet; and the Brahmaputra and Ganges, where the range is sixteen feet.[1] Since lakes, bays, gulfs, and inland seas have weaker waves and currents than the open sea, they are more favorable than the latter for the growth of deltas. Hence occur such deltas as those of the Mississippi, the Nile, the Po, and the Danube.

Deltas are likely to be absent, or confined to the heads of bays, on coasts which have recently sunk. Their general absence on the Atlantic coast of the United States is a case in point.

The following figures give some idea of the extent of deltas, and of their importance in land building. The Mississippi delta is advancing into the Gulf at the rate of about 100 yards per year, or a mile in 16 or 17 years. Its length is more than 200 miles, its area more than 12,000 square miles, and its depth at New Orleans has been estimated at 700[2] to 1000[3] feet. This great depth is believed to be the result of subsidence, and so of the superposition of one delta on another.[1] The delta of the Yukon has a sea margin of 70 miles, and extends more than 100 miles inland. The delta of the Rhône has also had a remarkable growth, considering the size and the history of the stream. Arles, near the debouchure of the stream, was 14 to 16 miles inland in the fourth century B.C., and is now 30 miles inland.[2] The Rhône has also built a great delta in Lake Geneva, and its lower delta is built of sediment gathered below the lake. The Po has built a delta 14 miles beyond Adria, the port which gave its name to the Adriatic Sea. The extension of this delta has been at the average rate of about 50 feet per year, but recently, on account of artificial embankments, the rate has been much more rapid.[3] The Ganges and Brahmaputra together have made a delta of great size. Its area is sometimes estimated to be as high as 50,000 or 60,000 square miles, and its head is more than 200 miles from the sea.[4] The head of the Nile delta is 90 miles from the sea, and it has a coastal border of 180 miles. The head of the delta of the Hoang-Ho is about 300 miles from the coast, and its seaward border has a length of about 400 miles, though with some highland interruptions.[5]

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[1] Davis. Physical Geography, p. 294.

[2] Humphreys and Abbot. Physics and Hydraulics of the Mississippi River.

[3] Corthell. Nat'l Geog. Mag., Vol. VIII, p. 351, 1897.

After a delta has been built into a lake, the lake may disappear, leaving the delta out of water. Such "fossil" deltas, if so recently exposed that erosion has not destroyed their distinctive features, are readily recognized by their flat tops, their abrupt and lobate fronts, and their characteristic structure. They are often a means of determining the former existence of extinct lakes,[6] or the former higher levels of lakes which still exist.[7] Elevated deltas on seashores show either a rise of the land or a depression of the sea-level.

The material which is carried along the coasts or shores from the mouths of rivers may take on various and peculiar forms, according to the strength, direction, and relations of waves and currents. The consideration of these forms belongs more properly to the work of the sea than to that of rivers, since rivers are not concerned in their construction except in supplying material.

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[1] Russell. Rivers of North America, p. 132.

[2] Prestwich. Chemical and Physical Geology, Vol. I, p. 85.

[3] Geike. Text-book of Geology, 3d ed., p. 402.

[4] Medlicott and Blanford, Geology of India, Chap. XVII; Medlicott, Records of the Geological Survey of India, 1881; Oldham, Geology of India, 2d ed., Chap. XVII; and Ferguson, Q. J. G. S., Vol. XIX, pp. 321–54. The extent of this and other deltas is variously stated, probably because it is difficult to determine the exact position of its head and borders.

[5] Dana. Manual of Geology, 4th ed., p. 198.

[6] Salisbury and Kümmel. Lake Passaic. Ann. Rept. of the State Geologist of New Jersey, 1893, and Jour. of Geol., Vol. III, p. 533.

[7] Gilbert. Lake Bonneville, Mono. I, U. S. Geol. Surv

Delta lakes. — Delta-building streams sometimes help to form lakes by throwing their deposits around an area which fails to be aggraded to sea-level. Lake Pontchartrain, and other lakes in the delta of the Mississippi are examples (Fig. 187).

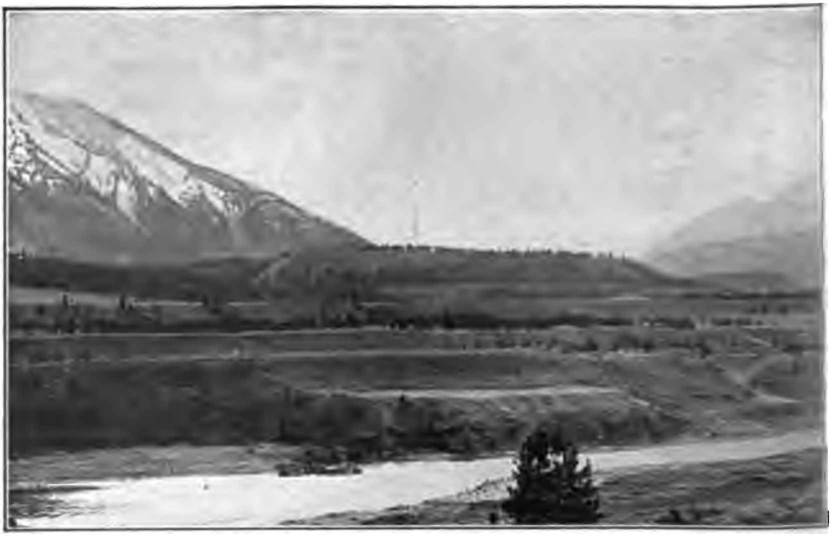

FIG. 192.—Terraces of the Frazier River at Lilloet, B. C. (Calvin.)

STREAM TERRACES.

Stream terraces[1] are bench-like flats or narrow plains along the si of valleys (Fig. 192). They are usually narrow, but sometimes h great length in the direction of the axis of the valley. They orig in various ways.

Due to inequalities of hardness. — Reference has already b made (p. 140) to the effect of hard horizontal layers in the develop of terraces and terraciform projections on the sides of valleys (Fig. 1 Such terraces are the result of differential degradation, and the up surface of the hard layer marks the lower limit of the terrace, which commonly has a distinct slope toward the stream. Except where interrupted by tributary valleys, such terraces are likely to be continuous in a valley so long as the structure remains the same and the stream sustains the same relation to it. Such terraces would first show themselves in the older part of the valley. The effect of inclination of the hard stratum on the development of such terraces will be readily inferred. Terraces and benches of this sort are not equally distinct at all stages of a valley's history. For great distinctness, the hard layer should have been exposed long enough to allow the general processes of erosion to have effected considerable differential wear, but not long enough to allow the topographic effects of unequal resistance to be obliterated.

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[1] For discussions of terraces see Gilbert's Henry Mountains, p. 126; D River Terraces in New England, Bull. of the Mus. of Comp. Zool., Geol. Series V, pp. 282-346; and Dodge, Proc Boston Soc. of Nat. Hist., Vol. XXVI, pp. 2

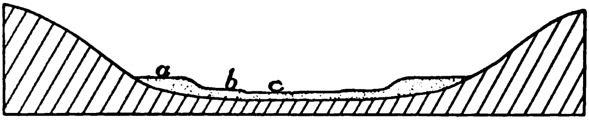

FIG. 193.—Diagram illustrating a distinct terrace and a "second bottom (*b*)," which may be regarded as a low terrace.

Normal flood-plain terraces. — It has been seen that deposition in a river valley stands in more or less definite relationship to the stage of its development, and that the deposition which leads to the development of an alluvial plain is likely to take place where the higher gradient of the upper course gives place to the gentler gradient of the lower. It has also been seen that as a stream's history advances, the stretch where the gradient is high recedes up-stream, and that the point which marks the head of active deposition follows. It follows that a river flat or flood-plain normally begins in the lower part of a valley, and works progressively headward, its upper end following, at some considerable distance, the head of the valley itself.

The commoner river terraces are remnants of former flood-plains, below which the streams which made them have cut their channels. It has already been pointed out (p. 184) that processes of erosion and deposition work together in the development of flood-plains, and that some flood-plains have but little alluvium (Fig. 174), while others owe their origin wholly to stream deposits (Fig. 175). It follows that terraces developed from flood-plains may be of rock, of alluvium, or of rock covered with alluvium.

The amount which a river channel must be deepened in order to change the remnants of its flood-plain to terraces cannot be definitely stated. When a channel is so deep that the remnants of a former flood-plain are no longer flooded, they would be called terraces, especially if a lower flood-plain has been developed. Even though not above the reach of floods, they are often called terraces if they are notably above the channel and separated from it by a lower plain. Thus the flat at *b*, Fig. 193, would be called a terrace, even though covered by water in exceptional floods; but the flat at *c*, but slightly above the channel, would hardly be called a terrace.

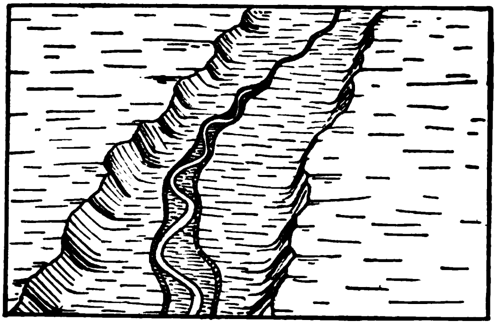

Fig. 194.—Diagram illustrating the beginning of the development of a terrace from a flood-plain.

The question now arises why a stream, having once developed a flood-plain, should sink its channel to a lower level, leaving parts of the old flood-plain as terraces. This may be brought about by the operation of various causes.

(1) In the first place, the head of the valley-plain where the first notable deposition takes place normally advances up-stream. After the advance has been considerable, the descending stream may, on reaching the head of its valley-plain, lose so much of its load as to be able to sink its channel into the flood-plain farther down the valley (Fig. 194).

(2) Ordinarily a stream does not drop all its load at the head of its plain, but only its excess; but it will always drop coarse sediment to take fine, if fine be available. For a relatively small amount of coarse material dropped, a relatively large amount of fine may be taken up (p. 179). Other things being equal, it follows that when a stream

drops coarse material to take fine, its channel is degraded unless there is at the same time a great reduction in the stream's energy. Such reduction is likely to go with the decreasing declivity down-stream; but this is partly, or sometimes wholly, counterbalanced by the increasing volume of water. By the exchange of load, therefore, a stream may ultimately sink its channel below the flood-plain which the earlier and perhaps smaller stream had developed.

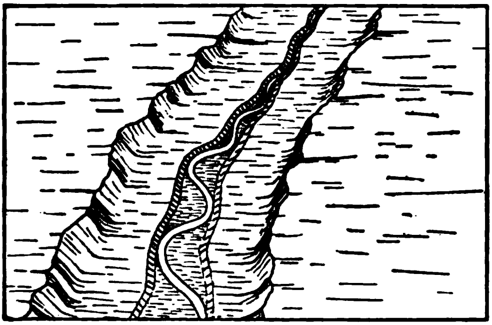

Fig. 195.—Diagram to illustrate the development of river terraces by the widening of a channel or meander belt. The valley flat above might not be called a terrace; but the same plain below, where the meander belt has some width, would be called a terrace.

(3) Again, so long as a stream is actively eroding at its head, there is likely to be some aggradation below. At a later stage in the stream's history, when active erosion at the head has ceased because of the reduction of the surface, less material will be carried from the upper part of the valley, and the stream on the flood-plain below, formerly loaded with material from up the valley, is now free to take up and carry away material temporarily left on the flood-plain. The result is a deepening of the channel.

(4) Any stream which has reached the flood-plain stage is likely to meander. After the flood-plain has become wide, the width of the belt within which the stream meanders is less than the width of its plain. In the Lower Mississippi, for example, the meander belt is often no more than a third to a tenth of the width of the flood-plain. It has already been pointed out that the meanders migrate down the valley. In so doing they depress the meander belt, the tendency being to reduce it to the level of the channel, and, therefore, below the level of the flood-

plain. As the meander belt widens, the depression which it develops becomes more and more capacious. Presently it may attain such dimensions as to hold the water of ordinary floods. At this stage, or even before, such parts of the earlier flood-plain as remain, are terraces.

These several tendencies conspire to partially destroy river flood-plains, and to transform such parts as remain into terraces *in the normal course of a river's history.* They appear first in the lower part of the valley, and migrate headward, following the course of nearly every other phase of activity in a stream's history. The heads of the terraces follow, at a respectful distance, the head of the flood-plain, just as the head of the flood-plain follows at a distance the head of the valley. The second and subsequent flood-plains and the terraces to which they give origin follow the same course.

Terraces developed by the normal activities of a stream are always low, and it is improbable that they would ordinarily be conspicuous. The vertical distance between the first (highest) and second is greater than that between the second and third. The principles developed on page 65 et seq., in connection with the erosion of the hypothetical island, are applicable here.

Flood-plain terraces due to other causes.—Certain other causes, accidental rather than normal to a stream, result in the development of terraces from flood-plains. (1) If there be uplift in a region where the rivers have flats, the streams are rejuvenated, and the remnants of their former flood-plains become terraces. (2) If an alluvial flood-plain has been built as the result of an excessive supply of sediment (p. 186), the exhaustion or withdrawal of the excessive supply would leave the stream again relatively clear, and free to erode where it had been depositing. It would forthwith set to work to carry away the material which it had temporarily unloaded on the plain. The plains built up in many valleys in the northern part of our continent during the glacial period, when the drainage from the ice coursed through them, have subsequently been partially destroyed by erosion, and their remnants have become terraces. A notable reduction in the amount of available sediment, even when the earlier supply was not excessive, produces a similar result. (3) A notable increase in the volume of a stream, without corresponding increase in load, as when one stream captures another, may occasion the development of terraces by allowing the stream to deepen its channel. (4) Above any barrier which dams a stream, a flood-plain is likely to be developed. When the barrier is removed the stream will cut more or less deeply into the plain above, leaving terraces. (5) The recession of a falls through a flood-plain converts such parts of it as remain, into terraces.

In conclusion, it may be stated that many river terraces, mostly very low, are normal features of valley development, coming into existence at definite stages in a valley's history. They are generally composed, in large part, of river alluvium. Others result from more or less accidental causes, working singly or in conjunction, and to this class belong all of the more conspicuous terraces developed from flood-plains. The structure of a terrace often affords some clue to its origin (Fig. 196).

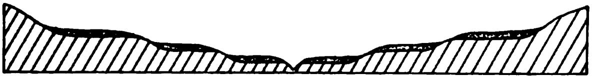

FIG. 196.—Terraces partly of rock and partly of alluvium. Such terraces indicate successive uplifts, or some other change which had a similar effect on the stream which made the valley.

Discontinuity of terraces.—When a stream sinks its channel into its flood-plain, it does not follow that a terrace remains on each side. Where the stream's deepened channel is in the middle of its flood-plain, there is, temporarily, a terrace on either side; but wherever the deepened channel is at one margin of its flood-plain, a terrace remains on the other side only. Even where continuous at the outset, terraces soon become discontinuous, for all processes of subaërial erosion conspire to destroy them. A stream is likely to meander on its second and later flood-plains, as on its first and highest one. Wherever the meanders on its second flood-plain reach the borders of the first flood-plain, the terrace at that point disappears, and since the meanders are continually migrating, terraces are continually disappearing. The same would be true of the second terrace, if a second were developed. The removal of portions of a terrace by the sweep of meanders is likely to leave the remnants cuspate toward the stream.[1] Again, tributary streams, in bringing their channels into topographic adjustment with their mains, cut through the terraces of the latter. New gullies develop on the faces of the terraces and their heads work back across them, dissecting them still further. At the same time, sheet erosion and other phases of slope wash tend to drive the scarps of the terraces back toward the bluff beyond. By the time a second series of terraces is well developed, no more than meagre remnants of the first may remain.

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[1] Davis, Bull. Mus. Comp. Zool., Geol. Ser., Vol. V.

From the foregoing considerations it is clear that the extent to which river terraces once developed, now remain, is dependent in part on the length of time which has elapsed since the river sank its channel below them. Other things being equal, the greater their age the more meagre their remnants.

Terraces developed from river plains formed chiefly by alluviation stand a better chance of long life than most other alluvial terraces. This is because of the configuration of the original valley, the aggradation of which gave origin to the plain. The principle involved is illustrated by Fig. 197. In developing the second flood-plain the river encounters the rock wall of the valley. This greatly retards lateral erosion, and the terrace above, *defended*[1] by the rock, is likely to be long-lived.

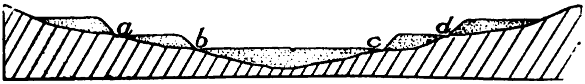

FIG. 197.—Diagram to show why certain terraces are longer lived than others.

Alluvial terraces, like rock shelves, are popularly thought to mark "old levels of the river." In one sense this is true, but not in the sense in which the expression is commonly used. Every level, from the crest of the bounding bluffs to the bottom of a valley, is a level at which water ran for a longer or shorter time; but the terrace does not mean that the river was once so much larger than now as to fill the valley from its present channel to the level of the terraces.

Termini of terraces.—From the mode of development of terraces it will be seen that, traced up-stream, each terrace should theoretically grade into a flood-plain at its upper end (Fig. 194), and that the upper end of the second (from the top) terrace, where there are two, would not be so far up-stream as the upper end of the first (highest). This is represented diagrammatically in Fig. 198.

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[1] This point has recently been emphasized by Davis, loc. cit., pp. 282–346.

The down-stream termini of terraces are rarely distinct. This is partly because the notable meandering of the streams in their lower courses is antagonistic to the preservation of terraces. If all terraces once developed remained, and if delta-building proceeded without interruption from waves, the relations should be somewhat as follows: Traced down-stream, the cliff between the oldest (highest) terrace and the next younger becomes gradually lower until it finally disappears, and the continuation of the two is found in a common plain. The cliff between the second and third terraces should disappear in the same way, and below its disappearance the plain representing their continuation is continuous with that representing the continuation of the first and second. The cliff between the second and third teraces may or may not continue farther down-stream than that between he first and second. The plains below the terraces finally become ontinuous with the lowest flood-plain and with the delta. These

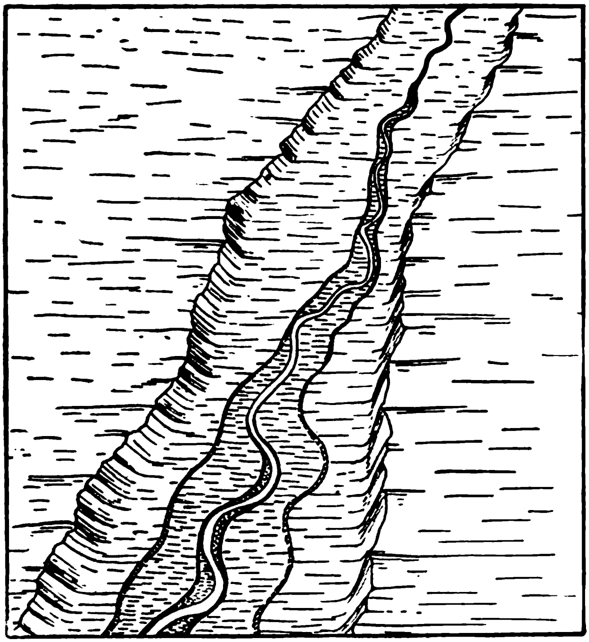


FIG. 198.—Diagram looking up the valley, showing two terraces below, one in the middle, and none above. The relations are purely diagrammatic.

relations can rarely be seen because of the destruction of the older terraces, and because of the erosion by waves along shore.

The topography of terraces is similar to that of flood-plains, except in so far as modified by erosion. While flat in general, the terrace may slope either toward or from the valley bluff, and its surface may be marked by all the minor irregularities which characterize a flood-plain.

CHAPTER IV.

THE WORK OF GROUND- (UNDERGROUND) WATER.

MANY familiar facts demonstrate the general presence of abundant water beneath the surface of the land. The thousands of wells in regions peopled by civilized races, and the countless springs which issue from the sides of mountains and valleys are a sufficient proof both of the wide distribution of ground-water and of its great abundance.

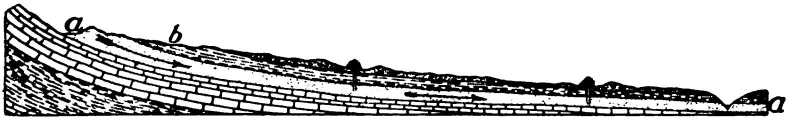
Certain well-known facts make it clear that ground-water is intimately connected with rainfall. In a dry season the level of the water in wells commonly sinks, and after a heavy rain it rises (p. 71); and the amount of sinking is greater when the drought is long, and the rise is most notable when the rainfall is heavy. Many springs which discharge large quantities of water during a wet season flow with reduced volume, or cease to flow altogether in periods of drought. Furthermore, the water of springs and wells has the properties which rain-water would possess after sinking beneath the surface and dissolving mineral substances. Rain-water is seen to sink beneath the surface with every shower, and since this source seems altogether adequate for ground-water, and since no other source is known whence any considerable amount of ground-water might come, it is concluded that atmospheric precipitation is its chief source.

Water gets beneath the surface by processes which are readily seen. Wherever the soil is porous some of the rain which falls upon it is absorbed. Sinking through the soil to the solid rock it finds cracks and pores through which it descends to great depths. Nowhere are the rocks beneath the mantle rock so compact and so free from cracks, when any considerable area is considered, as to prevent the percolation of water through them.

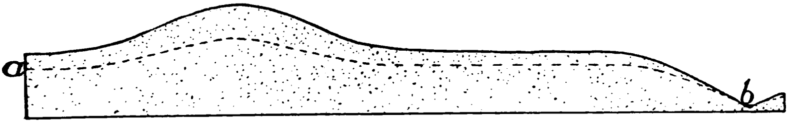
Conditions influencing descent of rain-water. — There are several conditions which influence not only the amount of water which sinks beneath the surface in a given area, but the proportion of the precipitation which follows this course. These are as follows: (1) *Amount of precipitation.*—In a general way it is true that the greater the amount of precipitation the greater the amount of water which will sink beneath the surface. Other things being equal, a region of heavy precipitation is a region where wells are easily obtained and springs common. (2) *Rate of precipitation.*—A given amount of precipitation may be concentrated in a short interval, or distributed through a considerable period of time. In the latter case more of the water sinks beneath the surface; in the former, a larger proportion runs off over the surface. The reason is readily seen. Water passes through small spaces, such as those of soil, slowly, and its rate of passage decreases rapidly with decreasing size of the passage-ways. When rain falls rapidly on a surface of even moderately close texture, the uppermost layer of soil is promptly filled with water, and since the water passes downward slowly, the uppermost saturated part of the soil becomes virtually impervious. While in this condition, the water which falls on it will run off if there be slope, and stand if there be none. In the latter case it will sink slowly as the water in the soil passes down to lower levels. If precipitation takes place no faster than the water can sink through the soil, all the water may become ground-water. (3) *The topography of the surface* has much to do with determining the proportion of rainfall which becomes ground-water. If the surface be flat, more will sink in; if it be sloping, more of it will run off before it has time to sink. Other things being equal, the steeper the slope the larger the proportion of the rainfall which will run off over it. (4) *The texture of the soil,* or other material on which the rain falls, helps to determine what proportion of it sinks beneath the surface. If the surface materials be porous, the water sinks readily; if of close texture, it finds less ready ingress. Other things being equal, the closer the texture of the soil the less the proportion of the rainfall which will enter it. (5) *The texture and structure of the rock* beneath the surface have some influence on the amount of ground-water. The rock may be stratified or massive; it may be abundantly or sparsely jointed; it may be porous or compact. On the whole, stratified rock is more favorable for the entrance of water than unstratified, partly because of its greater average porosity, and partly because the planes of division between beds often allow the passage of water. If the beds of stratified rock are vertical or inclined, water finds its way into them more readily than if they are horizontal, in so far as it descends along stratication planes. Horizontally bedded rock, or rock which is not bedded t all, may be so much jointed, and the joints so open, as to allow the ater to enter readily.

The conditions favorable to the sinking of abundant water below e surface are therefore heavy precipitation, falling slowly on a rface with little relief, a soil of open texture underlain by rock which porous, or affected by vertical or highly inclined planes of cleavage. he annual discharge of water by rivers is estimated to be about 22 per-nt. of the rainfall on the land.[1]

Supply of ground-water not altogether dependent on local rainfall. -The amount of ground-water in a given region is not always tirely dependent on the local rainfall. Ground-water is in constant ovement, and entering the soil or rock at one point it may, after a ng subterranean journey, reach a point far from that at which it tered. Thus beneath the Great Plains of the West there is much bterranean water which fell on the eastern slopes of the mountains the west. It has flowed beneath the surface to the Plains, where me of it is now withdrawn for the purposes of irrigation in regions ere rainfall is deficient. The accompanying diagram (Fig. 199) strates the flow here described.



. 199.—Diagram illustrating the general point that ground-water is not dependent entirely upon local supply.



. 200.—Diagram illustrating the position of the ground-water surface (the dotted line) in a region of undulatory topography.

The ground-water surface. Water table.—The water table has ady been defined (p. 71) as the upper surface of the ground-er. In a flat region of uniform structure the ground-water surface ssentially level, but rises and falls with the rainfall. Where the topography of a region is not flat, the ground-water surface is not level. As a rule it is higher, though farther below the surface, under an elevation than under surrounding lowlands, as illustrated by Fig. 200. The explanation is not far to seek. If a hill of sand be exposed to rainfall, most of the water falling on its porous surface will sink into it. If the precipitation continues long enough, as in a protracted rain, the hill of sand will be filled with water, the water occupying the interstices between the grains. If the sand of the hill could be removed, leaving the water which it contains on the same area, it would constitute a mound perhaps a third or a fourth as high as the hill itself. If unsupported, this mound of water would spread promptly in all directions until its surface was level. While the sand remains, the water in it constitutes a mound, and has a tendency to spread. It does in fact spread, but since the process involves great friction the spreading is slow. With the spreading the surface of the water in the sand sinks, and sinks fastest at the center where it is highest (*b*, Fig. 201). If the process were not interrupted the surface of the water in the hill would, in time, sink approximately to the level of the water in the surrounding land (*d*, Fig. 201); but at every stage preceding the last, the surface of the water would be higher beneath the summit of the hill than elsewhere, though farther from the surface. In regions of even moderate precipitation the water surface beneath the hills rarely sinks to the level of that in the lowlands adjacent, before being raised by further rains.

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[1] Murray. Scot. Geog. Mag., Vol. III, p. 70, 1887.

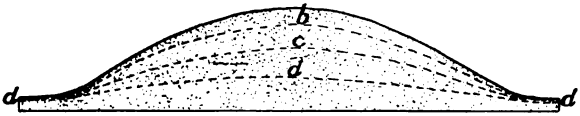

FIG. 201.—Diagram to illustrate the relations of ground-water to the surface.

The water-level beneath the lowlands also sinks. Some of it finds its way into valleys, some of it sinks to greater depths, and some of it evaporates; but since the water surface beneath the elevation sinks more rapidly than that beneath the lowland, the two approach a common level. Their difference will be least at the end of a drought, and greatest just after heavy rains.

Depth to which ground-water sinks.—The depth to which ground-water penetrates has not been determined empirically. No borings or excavations of any sort have been made to such depths as to indicate that its limit was being approached, though some of them are a mile or more deep. There is a popular belief that water sinks until it reaches a temperature sufficient to convert it into steam, but except for special localities where hot lava lies near the surface, this belief is not well founded. In the first place, it is not known at what temperature water below the surface would be converted into steam. While water boils at sea-level at a temperature of 212° (Fahr.) a higher temperature would be necessary below that level.

Assuming the temperature of water sinking beneath the surface to be 50° Fahr., its temperature must be raised 162° to bring it to the temperature at which it would boil at sea-level. On the above assumption of initial temperature, the following table shows the depths at which water would reach a temperature of 212° Fahr. under various assumptions as to the rate of increase of temperature. It shows also the pressure in atmospheres which would exist at these several depths if the overlying rock were full of water.

Rate of Increase of Temperature.	Depth at which Temperature of 212° would be reached.	Equivalent Pressure in Atmospheres.
1° for 50 feet..............	8,100 feet	238 (approximately)
1° for 60 "	9,720 "	285 "
1° for 70 "	11,340 "	333 "

With an initial temperature of 80°, corresponding to that of the warmer parts of the earth's surface, instead of 50°, the table would be as follows:

1° for 50 feet..............	6,600 feet	194 (approximately)
1° for 60 "	7,920 "	214 "
1° for 70 "	9,240 "	272 "

The temperature at which water boils increases with the pressure. A pressure of about 200 atmospheres is the critical pressure for water; that is, the pressure which, if increased, will prevent boiling altogether. The depth at which a pressure of 200 atmospheres would be reached, supposing the upper rock to be full of water, is about 6800 feet. The temperature of the water at this depth, under various assumptions as to initial temperature and rate of increase of heat, is shown in the following table:

Initial Temperature.	Rate of Increase of Temperature.	Temperature at a Depth of 6,800 Feet.
50°	1° for 50 feet	186° Fahr.
50°	1° for 60 "	163° "
50°	1° for 70 "	147° "
80°	1° for 50 "	216° "
80°	1° for 60 "	193° "
80°	1° for 70 "	177° "

Only one of these temperatures reaches the boiling-point of water *at sea-level*. It is therefore clear that at this depth water has not even closely approached the boiling temperature *for this depth*, and since this is the depth of the critical pressure, it is clear that it cannot boil at any greater depth. The descent of water is therefore not stopped, under normal conditions of crustal temperature, because it reaches its boiling-point. Locally, as in the vicinity of active or recently extinct volcanoes, the case may be different.

It is conceivable that water may descend until it reaches its critical temperature (somewhere between 610° and 635° Fahr.). The depth at which the critical temperature would be reached, under various assumptions, is shown in the following table:

Initial Temperature.	Rate of Increase of Temperature.	Depth of Critical Temperature.
50°	1° for 50 feet	28,000 to 29,250 feet
50°	1° for 60 "	33,600 to 35,100 "
50°	1° for 70 "	39,200 to 40,950 "
80°	1° for 50 "	26,500 to 27,750 "
80°	1° for 60 "	31,800 to 33,300 "
80°	1° for 70 "	37,100 to 38,850 "

There is good reason, in the increasing density beneath the surface, for believing that the rate of increase of temperature decreases with depth, and therefore that the rate of 1° for 50 feet for the depths concerned is too high. The greater depths of the table above are therefore believed to more nearly represent the truth than the lesser ones. (See discussion of underground temperature in Chapter XI.)

If descending water attained its critical temperature, the extent to which the resulting *water-gas* might be absorbed is not known. *So far as limited by temperature*, therefore, it is not possible to assign a limit to the descent of water under average conditions of crustal temperature.

Other considerations seem to place a limit to the descent of water. Rock, solid and unyielding as it seems, is yet plastic when under sufficiently great pressure. The cracks and cavities affecting it are believed to descend a distance which is but slight in comparison with the radius of the earth. Even if openings were once formed at greater depths, they could not persist, for the adjacent rock, under the pressure which there exists, would "flow" in, in effect (though perhaps not in principle) much as stiff liquid might, and close them. The outer zone of the earth where cavities may exist is known as the *zone of fracture.*[1] The depth of the zone of fracture differs for different rocks, but is not believed to extend below some such depth as five or six miles, even for the most resistant.[2] It is to be noted that these depths are less than those at which the critical temperature of water would be reached under most of the conditions, including all the more probable ones, specified in the above table.

Let it be assumed that water descends through openings in the rock to a depth of six miles. At this depth it would, under the various assumptions specified in the first and second columns of the following table, have the temperature indicated in the third column:

Initial Temperature.	Rate of Increase of Temperature.	Temperature at Depth of Six Miles.
50°	1° for 50 feet	683° Fahr.
50°	1° for 60 "	578° "
50°	1° for 70 "	502° "
80°	1° for 50 "	713° "
80°	1° for 60 "	608° "
80°	1° for 70 "	532° "

In two of these cases, namely, those in which the assumed rate of increase of temperature is highest, the temperature of the water at the assumed lower limit of the zone of fracture is above the critical temperature of water. If the assumptions involved in these two cases be correct, water might descend to the point where it would be converted into water-gas, and in this condition it might be occluded by the hot rock. In the other cases, involving the more probable assumptions, the critical temperature is not reached at a depth of six miles. If pores and cracks do not extend to greater depths, liquid water could not; and since the water at this depth has probably not reached its critical temperature, it cannot exist as water-gas. If it does not exist in the form of water-gas, its occlusion by the hot rock substance would not be probable. It would seem, therefore, that the descent of water under ordinary conditions is much more likely to be limited by the *zone of fracture*, than by temperature.

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[1] Hoskins. 16th Ann. Rept., U. S. Geol. Surv., p. 853.

[2] Van Hise. Principles of North American Pre-Cambrian Geology, 16th Ann. Rept., U. S. Geol. Surv.

Movement of ground-water.[1] — Ground-water is in more or less continual movement. If all the water be pumped out of a well it soon fills up again to its normal level by inflow from all sides. Springs and flowing wells also demonstrate the movement of ground-water. Near the surface the movement of ground-water is primarily downward if the medium through which it passes is equally permeable in all directions; but so soon as the descending water reaches the water surface, its descent is checked and its movement is partly lateral.

The commonest sort of movement of ground-water is that exemplified as the water sinks beneath the surface, namely, slow percolation through the pores and cracks of the soil and rock. Ground-water is not generally organized into definite streams, though underground streams, mostly small, are sometimes seen in caves and crevices, and sometimes issue as springs. Most underground streams which issue as springs probably have definite channels for short distances only before they issue. It is probable that ground-water frequently flows in considerable quantity along somewhat definite planes, without having open channels. Thus every porous bed of rock is likely to serve as the pathway along which subterranean drainage passes. This is especially true where the porous bed is underlain by an impervious one. The "reservoirs" from which artesian wells draw their supply are not usually streams or lakes, but porous beds of rock through which abundant water passes. As the supply is drawn off at one point, it is renewed by water entering elsewhere. Since the freedom of movement of ground-water is notably influenced by the porosity of the rock, and since the rock is, on the average, most porous and the pores largest near the surface, the movement of ground-water is, on the average, greatest near the surface, and least at its lower limit. In general the decrease of movement is much more rapid than the decrease in the size of the pores. It follows that while the upper part of the ground-water, especially that above ground-water level, moves somewhat freely, the lower part moves much more slowly. It is probable, indeed, that the movement in the lower part of the subterranean hydrosphere is extremely slight.

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[1] For a full discussion of this subject see King, 19th Ann. Rept., U. S. Geol. Surv., Pt. II, and Slichter, Water Supply and Irrigation, Paper No. 67, U. S. Geol. Surv.

The amount of ground-water. — The porosity of surface rocks varies widely, and the porosity of but few has been determined.[1] Such determinations as have been made are chiefly on building stones, in which the range of porosity varies from a fraction of one percent., in the case of granite, to nearly 30 percent. in the case of some sandstones. Building stone is perhaps more dense than the average surface rock. Furthermore, such tests as have been made do not take account of the larger cracks and openings of rock, for these would not appear in the specimens tested; nor of the mantle rock, which generally contains a large amount of water. From such determinations as have been made it is estimated that the average porosity of the outer part of the lithosphere is somewhere between 5 and 10 percent. If the porosity diminishes regularly to a depth of six miles, where it becomes zero, the average porosity to this depth would be half the surface porosity.[2] An average porosity of two and one-half percent. would mean that the rock contains enough water to form a layer nearly 800 feet deep. With an average porosity of 5 percent., this figure would be doubled.[3] While these figures are not to be regarded as measurements, they perhaps give some idea of the amount of ground-water. It is this sphere of ground-water which justifies the term *hydrosphere,* as applied to the waters of the earth.

Fate of ground-water.—Most of the water which sinks into the earth reaches the surface again after a longer or shorter journey. Some of it is evaporated from the surface directly; some of it is taken up by plants and is passed by them into the atmosphere; some of it issues in the form of springs; some of it seeps out; some of it is drawn out through wells; and much of the remainder finds its way underground to the sea or to lakes, issuing as springs beneath them. A small portion of the descending waters enters into permanent combination with mineral matter. Many minerals are known to take up water, being changed thereby from an anhydrous to a hydrous condition. It does not necessarily follow, however, that the total supply of water is thereby decreasing. Minerals once hydrated may be dehydrated subsequently, the water being set free. Furthermore, considerable quantities of water in the form of vapor issue from volcanoes, and volcanic vents often continue to steam long after volcanic action proper has ceased. It is probable that some, and perhaps much of the water issuing from these vents has never been at the surface before, and it is not now possible to affirm that the supply from this source does not offset, or even surpass, the depletion of the hydrosphere resulting from mineral hydration.

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[1] For tables see Buckley, Building and Ornamental Stones, Bull. IV, Wis. Surv., and Merrill, Stones for Building and Decoration, and various Survey Reports.

[2] It is probable that the porosity decreases in more than an arithmetic ratio, both because the deeper rocks are not of porous kinds, and because of the pressure which tends to close openings.

[3] Slichter (op. cit., p. 15) estimates that the ground-water is sufficient in amount to cover the earth's surface to a depth of 3000 to 3500 feet. Earlier estimates gave still higher figures (see Delesse, Bull. Soc. Geol., France, Second Series, Vol. XIX, 1861–62, p. 64).

THE WORK OF GROUND-WATER.

Ground-water effects very considerable results in the course of its history. These results are partly chemical and partly mechanical, the former being far more important than the latter.

Chemical Work.

The results of the chemical and chemico-physical action of water may be grouped in several more or less distinct categories.

1. The simplest effect is the *subtraction* of soluble mineral matter. Pure water is in itself a solvent of certain minerals; but the carbonic-acid gas extracted from the atmosphere, and the products of organic decay extracted from the soil make ground-water a much more efficient solvent. Something of the results which it achieves is shown by its composition. All ground-water, whether issuing as springs or drawn out through wells, contains much more mineral matter than the water which falls as rain, and the excess is acquired in its underground course.

The subtraction of soluble matter from rock renders it porous. The amount of material dissolved from a given place may be trivial or considerable, according to the character of the rock, the readiness with which water has access to it, and the character of the water. Locally, the subtraction of mineral matter may be the chief, or even the only appreciable, effect of the ground-water.

2. It sometimes happens that ground-water with certain mineral substances in solution exchanges them for other substances extracted

from the rock. Thus the process of *substitution* is effected. By this process the lime carbonate of a shell imbedded in rock may be removed, molecule by molecule, and some other substance, such as silica, left in its place. When the process is complete, the substance of the shell has been completely removed, though its form and structure are still preserved in the new material which has taken the place of the old. Buried logs are sometimes converted into stone by the substitution of mineral matter for the vegetable tissue. This is *petrification*. Petrification is altogether distinct from *incrustation*, which simply means the coating of an object with mineral matter. A bird's nest may be incrusted with lime carbonate, but it is not thereby petrified. Solution is a necessary antecedent of substitution.

3. The materials which are subtracted from the rock at one point may be added to other rock elsewhere. Thus a third type of change, *addition*, is effected. Rock may at one time and place be rendered porous by the subtraction of some of its substance, and the openings thus formed may subsequently become the receptacles of deposits from solution. This is exemplified in the stalactitic deposits of many caves. Not uncommonly cracks and fissures are filled with mineral matter deposited by the waters which pass through them. Thus arise *veins* which, for the most part, are nothing more than cracks and crevices filled by mineral matter brought to them in solution, and precipitated on their walls. Most veins of metallic ores have originated in this way.

4. A further series of changes is effected by ground-water when it, or the mineral matter it contains, enters into combination with the mineral matter through which it passes. One of the commonest processes of this sort, hydration, has already been referred to (pp. 43 and 428); but in the development of many of the commoner hydrous minerals changes other than hydration are involved. These changes result in *new mineral combinations*, the new minerals being developed out of the old, usually with some additions or subtractions. In the long course of time changes of this sort may be very great, so great indeed that large bodies of rock are radically changed, both chemically and physically. Much of the old substance may remain, but it has entered into new and more stable combinations with the materials which the water has brought to it.

Quantitative importance of solution. — In general, solution is probably most effective at a relatively slight distance below the surface. In the outermost zone of mantle rock the materials are usually less soluble than below, for they often represent the residuum after the soluble parts of the formation from which they originated were dissolved out. Below this zone the rock contains more soluble matter, and the water, charged with organic matter in its descent through the soil, is in condition to dissolve it. At greater depths the water has become saturated to some extent, and, so far forth, less active. Here, too, the movement is less free. The increased pressure at considerable depths, on the other hand, facilitates solution, which must be understood to take place under proper circumstances in any zone reached by the water.

Calculations have been made which illustrate the quantitative importance of the solution effected by ground-water. The springs of Leuk (Switzerland) bring to the surface annually more than 2000 tons of calcium sulphate (gypsum) in solution, and in the same time the springs of Bath (England) bring up an amount of mineral matter in solution sufficient to make a column 9 feet in diameter, and 140 feet high.[1]

The amount of mineral matter in solution in streams is also significant, for while stream-water is not all derived from ground-water, much of it had such an origin. In the case of several streams, among them the Thames and the Elbe, careful estimates of the amount of dissolved mineral matter have been made. Though the Thames drains an area only about one-tenth as large as the State of New York, it is estimated to carry about 1500 tons of mineral matter in solution to the sea daily.[2]

From the uppermost 20,000 square miles of its drainage basin the Elbe is estimated to carry yearly about 1,370,000 tons of mineral matter in solution. Estimates of the amounts of material carried to the sea in solution by several rivers are given on pp. 102 and 103. Much of this matter was brought to the rivers by waters which had been underground before reaching the streams.

From these figures it is clear that we have to reckon here with a very considerable factor in the lowering of land surfaces. From the amount of lime carbonate carried by the Thames it has been estimated that the average amount of this material dissolved from the limestone area drained by this stream is 143[1] tons per square mile per year. It is estimated that, on the average, something like one-third as much matter is carried to the sea in solution as in the form of sediment, and that by this process alone land areas would be lowered something like one foot in 13,000 years.[2]

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[1] Geikie. Text-book of Geology, 3d ed., p. 367.

[2] Ibid., p. 378.

Deposition of mineral matter from solution.—The deposition of material from solution is effected in several ways. (1) It is sometimes deposited by evaporation. This is well shown where water seeps out on arid lands. The same process is illustrated when water is boiled. (2) Reduction of temperature often occasions deposition. In general, hot water is a better solvent of mineral matter than cold,[3] and if it issues with abundant mineral matter in solution the precipitation of some of it is likely to take place. (3) Plants sometimes cause the precipitation of mineral matter from solution. About some hot springs, even where the temperature of the water is very high small plants of low type (algæ) grow in profusion. In ways which are not perfectly understood these algæ extract the mineral matter from the hot water. They are now thought to be a chief factor in the deposits about the hot springs of the Yellowstone Park.[4] The influence of organisms on precipitation from solution is not confined to the waters of hot springs. (4) A fourth factor involved in the deposition of mineral matter from solution is pressure. Pressure increases the solvent power of water with respect to minerals directly; it produces the same effect indirectly by its effect on the solution of gases. As water charged with gas comes to the surface, the pressure is relieved and some of the gas escapes. Such mineral matter as was held in solution by the help of the gas which escapes is then precipitated. (5) Precipitation is also sometimes effected by the mingling of waters containing different mineral substances in solution. Such mingling of solutions would be most common along lines of ready subterranean flow, and while each portion of the water entering a crevice or porous bed may be able to keep its own mineral matter in solution, their mingling may involve chemical changes resulting in the formation of insoluble compounds, and therefore in deposition. This principle has probably been involved in the filling of many fissures and crevices, converting them into veins. (6) The escape of gases from water, whether from increase of temperature or by the disturbance of water, sometimes causes the deposition of mineral matter held in solution.

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[1] Prestwich, Q. J. Geol. Soc., Vol. XXVIII, p. lxvii.

[2] Reade. Liverpool Geol. Soc., 1876 and 1884.

[3] This is not true in the case of minerals, such as lime carbonate, dissolved under the influence of gases in solution in the water.

[4] Weed. The Formation of Hot Springs Deposits. Excursion to the Rocky Mountains. Compte Rendu. Fifth Session of the International Geological Congress, p. 360, and Ninth Ann. Rept. U. S. Geol. Surv., pp. 613-76. Also B. M. Davis, Science, Vol. VI, pp. 145–57, 1897.

The deposition of material held in solution is most notable at two zones, one below that of most active solution, and the other at the surface, where evaporation is active. Under proper conditions, however, deposition may take place at any level reached by water.

Mechanical Work.

The mechanical work of ground-water is relatively unimportant. Wherever it is organized into definite streams, the channels through which it flows are likely to be increased by mechanical erosion as well as by solution. Either beneath the surface, or after the streams issue, the mechanical sediment carried will be deposited.

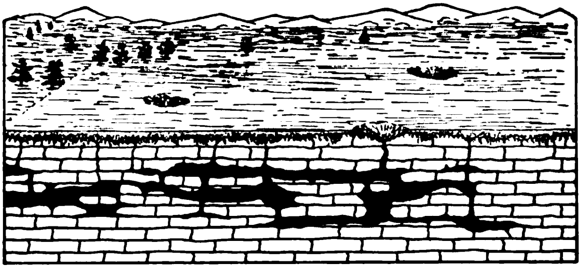

FIG. 202.—Diagram to illustrate the general form and relations of caves developed by solution. The black portions represent the cavern spaces. Some limestone sinks are represented on the surface above.

RESULTS OF THE WORK OF GROUND-WATER.

Weathering.—Where the solution effected by ground-water in any locality is slight and equally distributed, the result is to make the rock porous. If, for example, some of the cement of sandstone is dissolved, the texture of the rock becomes more open; but if all the cement be removed the rock is changed from sandstone to sand. If a complex crystalline rock contains among its many minerals some one which is more soluble than the others, that one may be dissolved. This has the effect of breaking up the rock, since each mineral acts as a binder for the rest. It might happen that no one of the minerals is dissolved completely, but that some one of them is decomposed by water, and certain of its constituents removed. Such change would be likely to cause the mineral so affected to crumble, and with its crumbling, if it be an important constituent of the rock, the integrity of the rock is destroyed. Where considerable chemical changes, especially subtractions, are going on, the rock is likely to crumble. The increase in volume attendant on hydration, etc., sometimes leads to the disruption of rock. These are phases of weathering. (For other phases of weathering see pp. 54 and 110.)

Caverns.[1]—Where local solution is very great results of another sort may be effected. In formations like limestone, which are relatively soluble, considerable quantities of material are frequently dissolved from a given place. Instead of making the rock porous, in the usual sense of the term, large caverns may be developed (Fig. 202). In their production, solution may be abetted by the mechanical action of the water passing through the openings which solution has developed. Considerable caves are found chiefly in limestone. They were probably developed when the surface relief was slight, and surface drainage therefore poor. Regions where caves were developed under these conditions may subsequently acquire relief, so that caves are not now confined to flat regions.

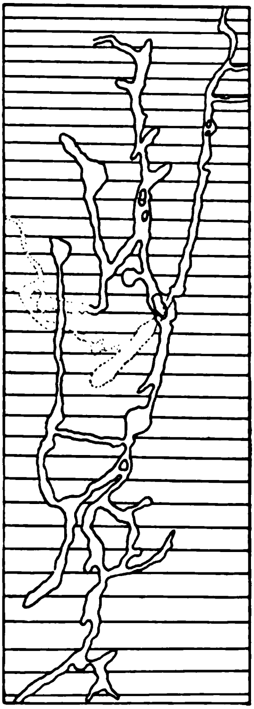

FIG. 203.—Ground-plan of Wyandotte Cave. The unshaded areas represent the passageways. (21st Ann. Rept., Ind. Geol. Surv.)

One of the best known regions of caves is in the basin of the Ohio in Kentucky and southern Indiana, where the number of caves is large, and the size of some of them, such as Mammoth and Wyandotte,

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[1] For a racy and interesting account of caverns see Shaler's Apects of the Earth

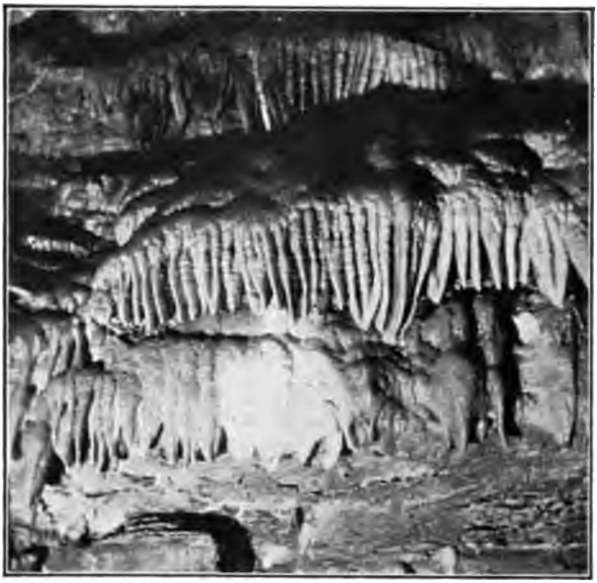

Fig. 204.—Deposits in Wyandotte (Ind.) Cave. (Hains.)

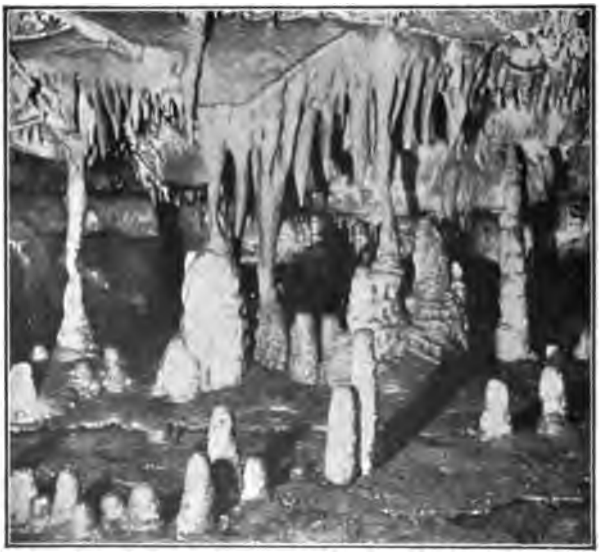

Fig. 205.—Deposits in Wyandotte Cave. (Hains.)

very great. A ground-plan of Wyandotte (Ind.) Cave is shown in Fig. 203. The aggregate length of the passageways is about 23½ miles.

Deposition often takes place in caves after they are formed (Figs. 204 and 205). It may even go on at the same time that the cave is being excavated. Here are formed the well-known stalactites and stalagmites. A stalactite may start from a drop of water leaking through the roof of the cave. Evaporation, or the escape of some of the carbonic gas in solution, results in the deposition of some of the lime carbonate about the margin of the drop, in the form of a ring. Successive drops make successive deposits on the lower edge of the ring, which grows downward into a hollow tube through which descending water passes, making its chief deposits at the end. Deposition in the tube may ultimately close it, while deposition on the outside, due to water trickling down in that position, may greatly enlarge it.

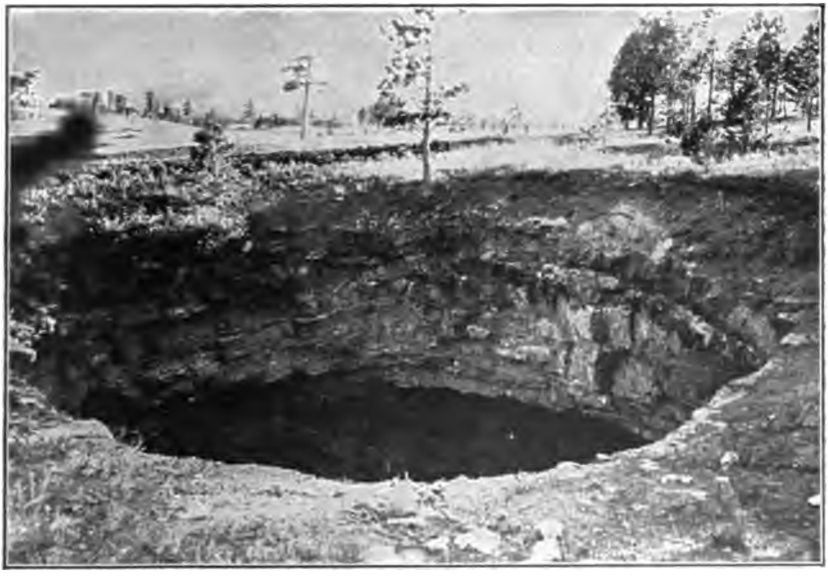

Fig. 206.—A limestone sink-hole, east-northeast of Cambria, Wyo., exceptional for its steep sides. Minnekahta limestone. (Darton, U. S. Geol. Surv.)

Underground caves sometimes give rise to topographic features which are of local importance. When the solution of material in a cavern has gone so far that its roof becomes thin and weak, it may collapse, giving rise to a sink or depression in the surface over the site

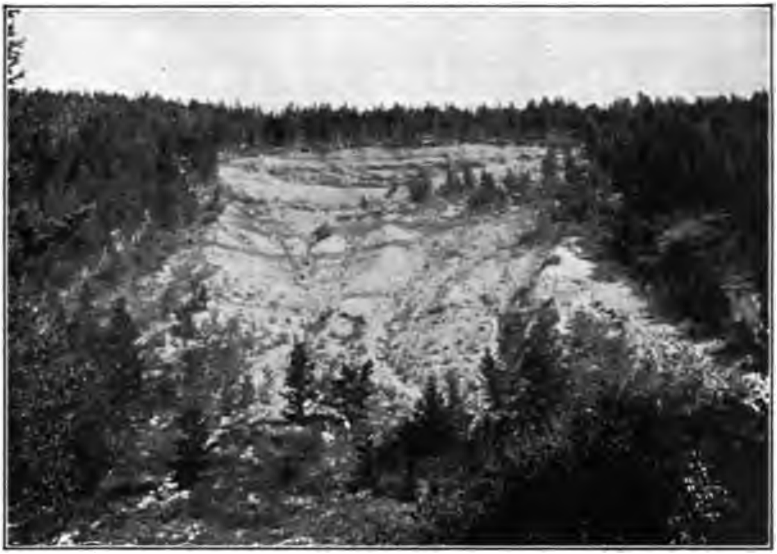

Fig. 207.—A fresh landslide near Medicine Lake, Mont. The bare space shows the position from which the slide started. (Whitney.)

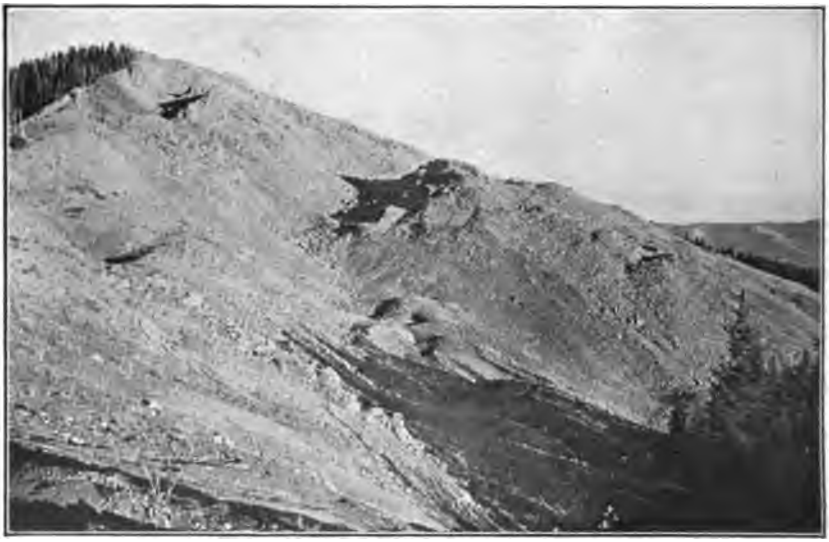

Fig. 208.—Landslide topography. The protruding mass on the right has slumped down from the mountain to the left. South face of Landslip Mountain, Colo., seen from the west; Rico quadrangle. (Cross, U. S. Geol. Surv.)

of the original cave. This is so common that regions of limestone caves are often affected by frequent sinks formed in this way. They are a conspicuous feature of the landscape in the cave region of Kentucky, and are well known in many other limestone districts. They are known as *limestone sinks*. (Fig. 206 and Fig. 2, Pl. XVII.)

Under certain circumstances caves may give rise to striking features of another sort. If for any reason the roof is destroyed at the two ends of a cave, remaining intact over the middle, the latter part constitutes a *natural bridge*. Natural bridges also originate in other ways (pp. 151, 153).

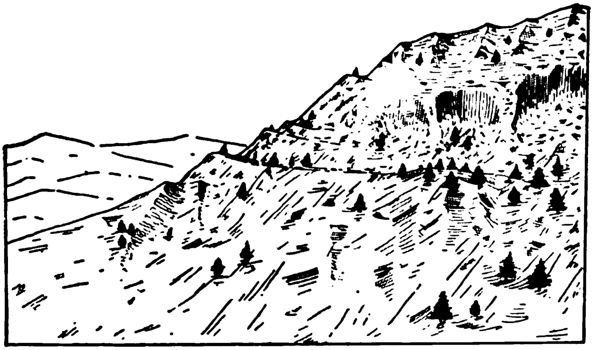

FIG. 209.—Landslide topography on Badger Mountain, Washington. The slumping material in this case is basalt.

Creep, slumps, and landslides.—When the soil and subsoil on a slope become charged with water they tend to move downward. When the movement is too slow to be sensible it is called *creep*. The common downward inclination of trees growing in such situations, the result of the more rapid creep of the surface as compared with the deeper part of the soil, is both an expression of the movement and of its slowness. Other factors besides ground-water are involved in creep (see p. 112).

When the movement is rapid enough to be sensible the material is said to *slump* or *slide*. This may happen when the slope on which water-charged mantle rock lies is steep (Fig. 207). Great landslides of this sort have been recorded, and some of them have done great damage. Where a stream's banks are high, and of unindurated material, such as clay, considerable masses sometimes slump from the bank or bluff into the river, or settle away slowly from their former positions. This is a common phenomenon along streams which have cut valleys in drift, and along shores on which waves are encroaching. The same phenomenon is common on a larger scale on the slopes of steep mountains.[1] Considerable terraces are sometimes developed on their slopes in this way, but they are usually irregular and discontinuous (Figs. 208, and 209). The loose débris on steep slopes sometimes assumes a sort of flowing motion and descends the slope with some such form and at some such rate as a glacier. Such bodies of débris are sometimes called "talus glaciers" (Fig. 210). In many such cases, snow and ice have had some part in their development.

In creep and in landslides gravity is the force involved, and the ground-water only a condition which makes gravity effective. Gravity alone accomplishes similar results, as illustrated by Fig. 211.

Summary.

All in all, ground-water is to be looked upon as a most important geological agent. When it is remembered that a very large part of all the water which falls on the surface of the earth, either in the form of rain or snow, sinks beneath the surface; that much of it sinks to a great depth; that much of it has a long underground course before it reappears at the surface; that it is everywhere and always active, either in subtracting from the rock through which it passes, in adding to it, in effecting the substitution of one mineral substance for another, or in bringing about new chemical combinations; and when it is remembered that this process has been going on for untold millions of years, it will be seen that the total result accomplished must be stupendous. The rock formations of the earth to the depths to which ground-water penetrates are to be looked upon as a sort of chemical laboratory through which waters are circulating in all directions, charged with all sorts of mineral substances. Some of the substances in solution are deposited beneath the surface, and some are brought to the surface where the waters issue. Much of the material brought to the surface in solution is carried to the sea and utilized by marine organisms in the making of shells. Without the mineral matter brought to the sea by springs and rivers, many shell-bearing animals of great

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[1] Russell has emphasized this point in 20th Ann. U. S. Geol. Surv., Pt. II, pp. 193–202, and Cross, 21st Ann. U. S. Geol. Surv., Part II, pp. 129–150.

PLATE XVII.

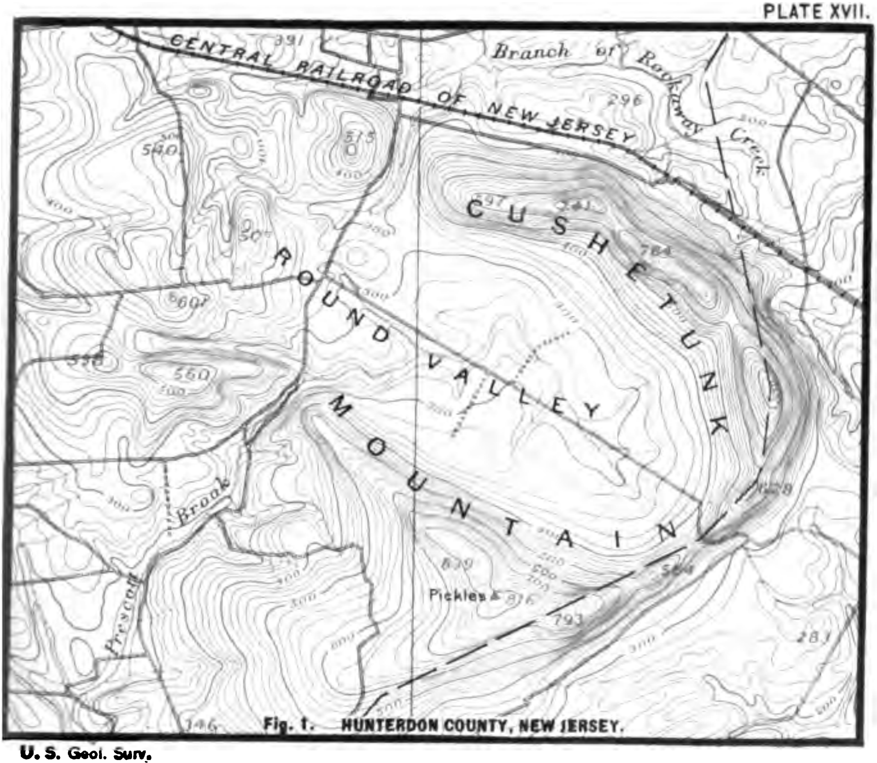


Fig. 1. HUNTERDON COUNTY, NEW JERSEY.

U. S. Geol. Surv.

Scale, 1+ mile per inch.

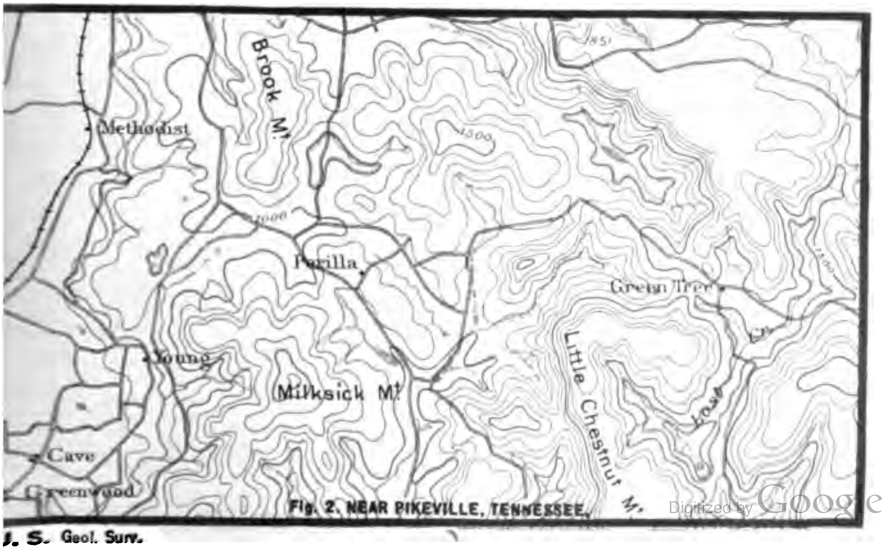


Fig. 2. NEAR PIKEVILLE, TENNESSEE.

U. S. Geol. Surv.

Scale. 2+ miles per inch.

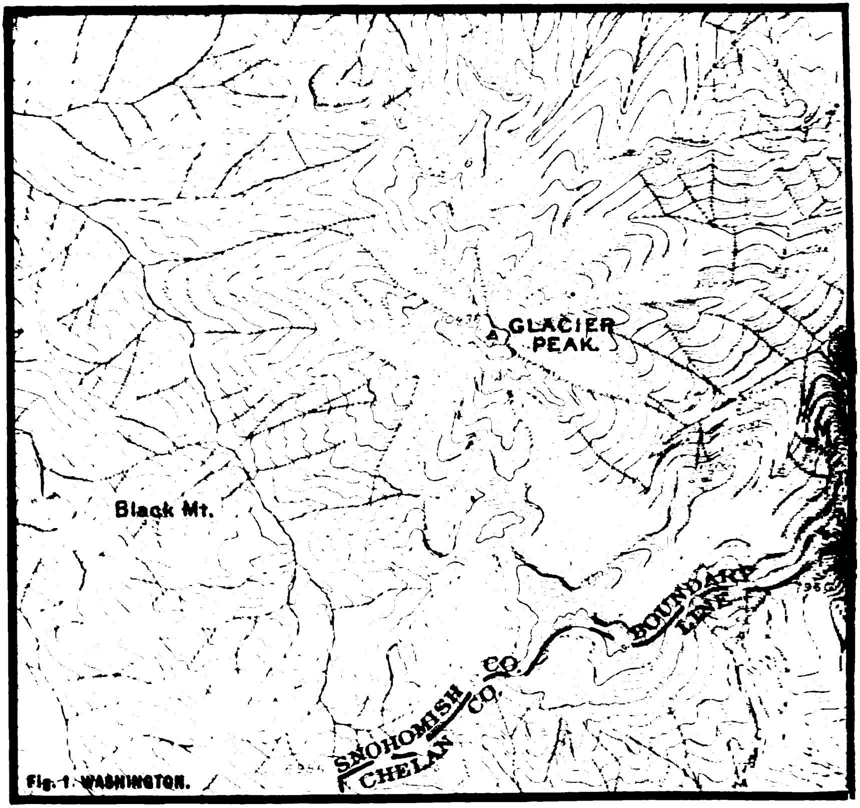
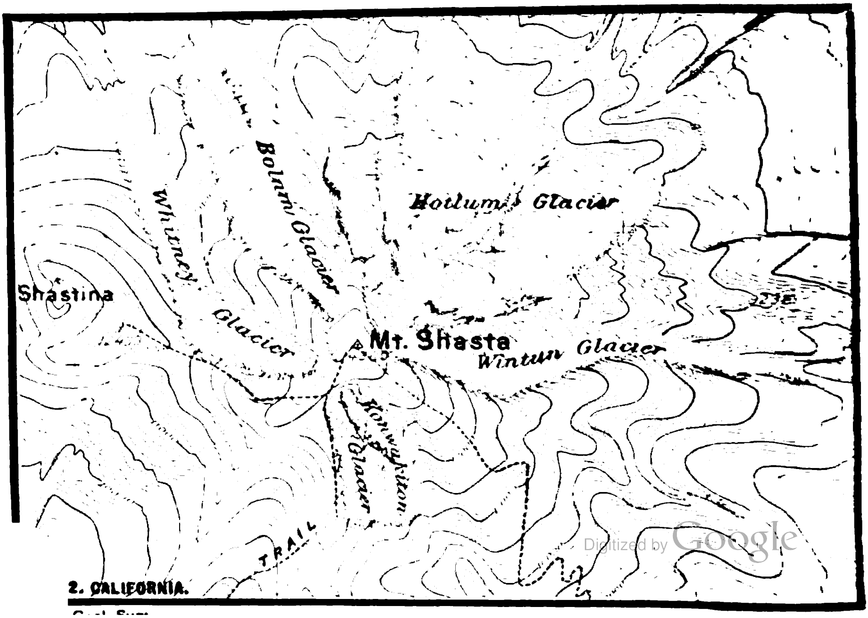


Fig. 1. WASHINGTON.

U. S. Geol. Surv.

Scale, 2+ miles per inch.



2. CALIFORNIA.

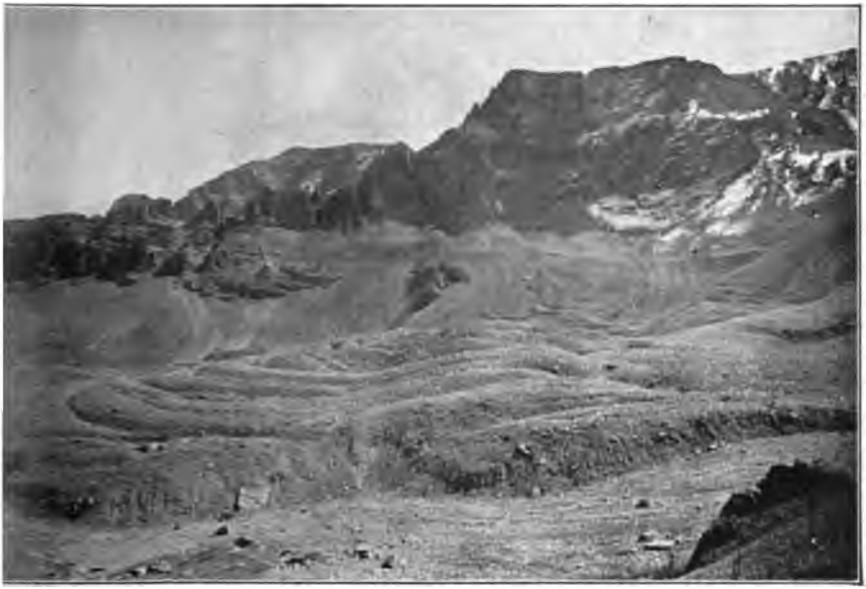
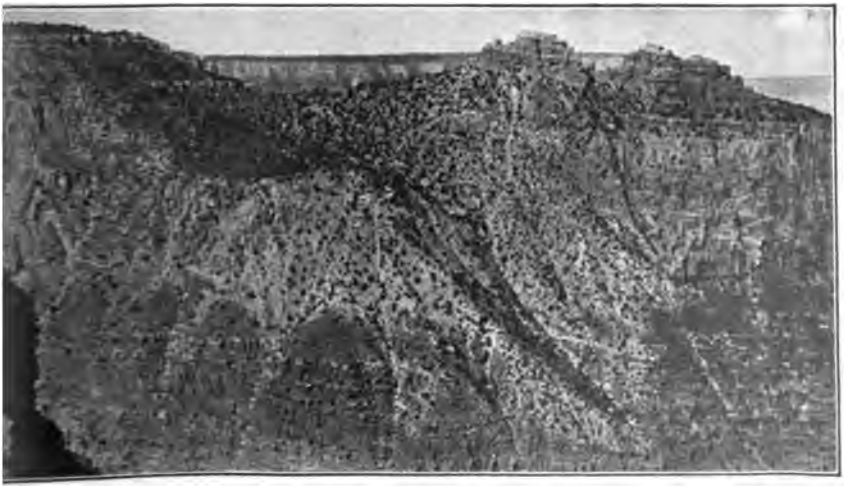


FIG. 210.—A "talus glacier." in Silver Basin, near Silverton, Colo. (Cross, U. S. Geol. Surv.)



211.—A huge mass of rock settling into the Canyon of the Colorado. A result of gravity action. (Atwood.)

importance, geologically, would perish. Biologically, therefore, as well as geologically, ground-water is of great importance.

SPRINGS AND FLOWING WELLS.

The term *spring* is applied to any water which issues from beneath the surface with sufficient volume to cause a distinct current. If the water issues so slowly as to merely keep the surface moist, it is not called a spring, but seepage. The spring from which water issues with a strong current, especially if it be upward, is comparable to a flowing well, while the spring from which water issues with little force, and without upward movement, is comparable to the flow of water into a common well.

Springs often issue from the sides of valleys (Fig. 212), the bottoms of which are below ground-water level. They are especially likely to issue at the surface of relatively impervious layers, and where the valley slopes cut joints, porous beds, or other structures which allow free flow of ground-water.

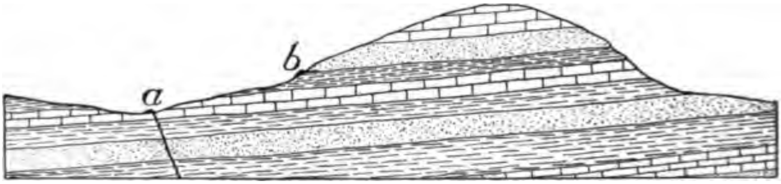


Fig. 212.—Diagram illustrating positions, *a* and *b*, favorable for springs.

Springs are classified in various ways, and these several classifications suggest characteristics worthy of note. They are sometimes said to be *deep* and *shallow*. The "deep" spring, as the term is ordinarily used, is one which issues with great force, and with something of upward movement, and the "shallow" spring, one which issues with little force, and without upward movement; but the spring which issues with force is not necessarily deep, nor is the one which issues with little force necessarily shallow. The idea involved in this grouping would be better expressed by *strong* and *feeble*. Springs are also classified as *cold* and *thermal*, the latter term meaning simply that the temperature is such as to make the springs seem warm or hot. The temperature of thermal springs ranges up to the boiling-point of water.

Between deep springs and shallow ones, and between cold springs and thermal, respectively, there is no sharp line of demarkation. Again, some springs are continuous in their flow, while others are *intermittent*. Most intermittent springs flow after periods of precipitation, but dry up during droughts (see p. 202). Springs are also classified as *mineral* and *common*. Mineral springs, in the popular sense of the term, are of two types: (1) Those which contain an unusual amount of mineral matter, and (2) those which contain some unusual mineral. Springs are especially likely to be called mineral if the substances which they contain, have, or are supposed to have, some medicinal property. All springs which are not "mineral" are "common." This classification is not altogether rational, for all springs contain more or less mineral matter, and many springs which are "common," contain more mineral matter than some springs that are "mineral." Mineral springs are themselves classified according to the kind and amount of mineral matter they contain. Thus *saline* springs contain salt; *sulphur* springs contain compounds (especially gaseous) of sulphur; *chalybeate* springs contain iron compounds, especially the sulphate; *calcareous* springs contain abundant lime carbonate, etc. These various mineral substances are extracted from the rock, sometimes by simple solution, and sometimes by solution resulting from other chemical change. The salt of saline springs is usually extracted from beds of salt beneath the surface. Lime carbonate, one of the commonest substances in solution in ground-water, is dissolved from limestone, or derived by chemical change from rocks containing other calcium compounds. Thus lime feldspars, by carbonation, give rise to lime carbonate. The chalybeate waters often arise from the oxidation of iron sulphide, a mineral which is common in many sedimentary rocks. The iron sulphate is itself subject to change in the presence of the ubiquitous lime carbonate. From this change iron carbonate results, and this is usually quickly altered to iron oxide, which, being relatively insoluble, is precipitated. About chalybeate springs, therefore, iron oxide is frequently being deposited. *Medicinal* springs are those which contain some substance or substances which have, or are supposed to have, curative properties.

Mineral matter in solution.—The number and variety of mineral substances in spring water is very great, and the amount of solid matter in solution varies widely. Some of the hot springs of the Yellowstone Park contain nearly three grams (2.8733) of mineral matter per kilogram.[1]

The composition of various spring and well waters is shown in the accompanying table, which gives some idea of the range of mineral substances commonly in solution in ground-water.

Geysers.—Geysers are intermittently eruptive hot springs. They occur only in volcanic regions (past or present) and in but few of them. Active geysers are virtually confined to the Yellowstone Park and Iceland, though they formerly existed at other places. Those of New Zealand have but recently become extinct. The great geyser region of the world is the Yellowstone Park, where there are said to be more than sixty active geysers.

The cause of the eruption is steam. The surface-water sinks down until, at some unknown depth, it comes into contact with rock sufficiently hot to boil it. The source of the heat is not open to inspection, but it is believed to be the uncooled part of an extrusive lava flow, or of an intrusive lava mass. From what was said on pp. 216 and 217 it is clear that geysers do not have their origin in water which sinks down to the zone of great heat, where the increment of heat is normal.

The water of a geyser issues through a tube of unknown length. Whether the tube is open down to the source of the heat is not determinable, but water from such a source finds its way to the tube. Water may enter the tube from all sides and at various levels from top to bottom. The heating may precede or follow its entrance into the tube, or both. So far as the water is heated after it enters the tube, the point of most rapid heating may be at the bottom of the tube or at some point above. If the temperature of the source of heat were high enough to convert the descending water into steam as fast as it enters the tube, the steam would escape continuously, though there would be no geyser; but if the rock is only hot enough to bring the water to the boiling-point after some lapse of time, and after some water has accumulated, an eruption is possible.

The exact sequence of events which leads up to an eruption is not known, but a definite conception of the principles involved may perhaps be secured by a definite case. Suppose a geyser-tube filled with water, and heated at its lower end. As the water is heated below, con-

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[1] Gooch and Whitfield. Bull. 47, U. S Geol. Surv.

Waters..........ring	Hot Spring	Boiling Spring	Warm Spring
Location.........tation, R.	Ward's Ranch, base of Granite Mts., Nev.	Shaffer's Ranch, Honey Lake Valley, Cal.	Warm Spring Sta. B. & B. R. Mono Basin
Date...............	....	....	....
Analyst..........ard	T. M. Chatard	T. M. Chatard	T. M. Chatard
References.......	*Ante*, p. 53	*Ante*, p. 51	Bulletin No. 9, U. S. Geol. Survey, p. 27
Sodium, Na....	.3554	.3040	.6116
Potassium, K...	.0191	.0094	.0630
Calcium, Ca....	.0367	.0121	.0589
Magnesium, Mg.	.0034	.0004	.0604
Barium, Ba....			
Strontium, Sr...			
Lithium, Li.....			
Iron, Fe.........			
Manganese, Mn..			
Chlorine, Cl.....	.2396	.2070	.2272
Bromine, Br.....			
Iodine, I.........			
Fluorine, Fl.....			
Carbonic acid, C	Trace	......	.5787
Sulphuric acid, H	.3901	.3492	.3131
Phosphoric acid,			
Boracic acid, H_3			
Alumina, Al_2O_3..	......	......	.0018
Silica, SiO_2......	.1136	.1310	.1220
Hydrogen in bica			
Organic substance			
Oxygen, O......[3]	.0255[3]	.0080[3]	.0325[3]
	1.1834	1.0211	2.0692
Carbonic acid, CO	......	Trace	
Sulphuretted hyd			

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[1] Correction fo as specific gravi

[3] Oxygen added to SiO_2 to form SiO_3 of Na_2SiO_3.

[2] Liters of gas thrown off per liter of water.

vection tends to distribute the heat throughout the column of water above. If convection were free, and the tube short, the result would be a boiling spring; but if the tube is long, and especially if convection is impeded, the water at some level below the surface may be brought to the boiling-point earlier than that at the top. Under these circumstances if even a little water in the lower part of the tube is converted into steam, the steam will raise the column of water above, and it will overflow. The overflow relieves the pressure on all parts of the column of water below the surface. If before the overflow there was any considerable volume of water essentially ready to boil, the relief of pressure following the overflow might allow it to be converted into steam suddenly, and the sudden conversion of any considerable quantity of water into steam would cause the eruption of all the water above it (Figs. 213 and 214). The height to which the water would be thrown would depend upon the amount of steam, the size and straightness of the tube, etc.

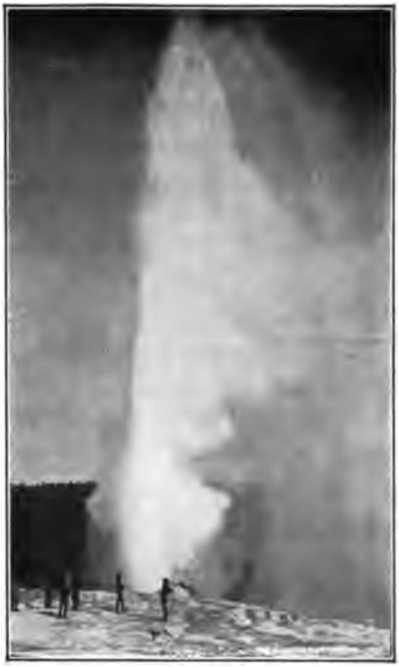
FIG. 213.—"Old Faithful" in eruption.

It is clear that everything which impedes convection in the geyser tube will hasten the period of eruption, since impeded circulation will have the effect of holding the heat down, and so of bringing the water at some level below the top more quickly to the boiling-point. It follows that anything which chokes up the tube, or which increases the viscosity of the water, or its surface tension, would hasten an eruption.[1]

Geysers often build up crater-like basins or cones (Figs. 214 to 217)

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[1] Weed. Ninth Ann. Rept. U. S. Geol. Surv., pp. 613–76, and Am. Jour. Sci., Vol. XXXVII, 1889, pp. 351–59.

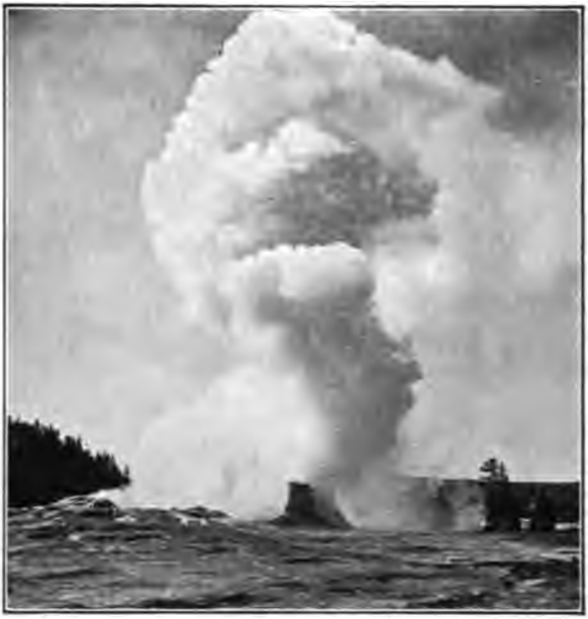

FIG. 214.—"Giant" Geyser, Yellowstone Park, in eruption. Shows also the cone. (Wineman.)

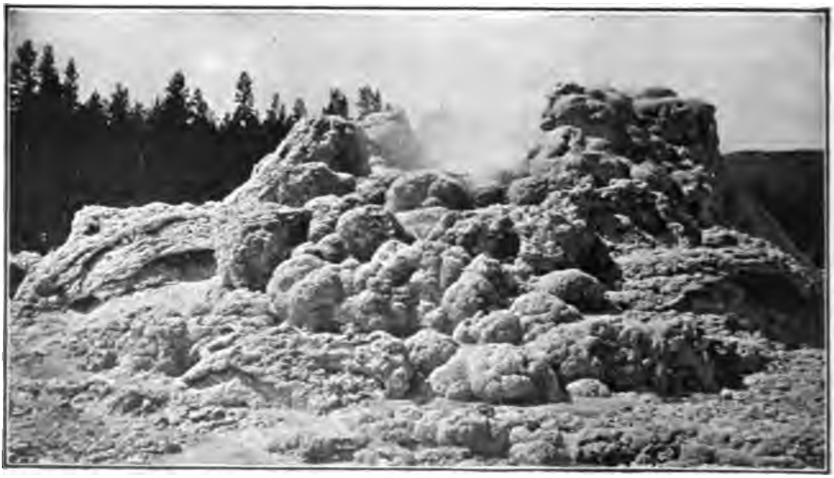

FIG. 215.—Cone (or crater) of Castle Geyser, Yellowstone Park. (Detroit Photo. Co.)

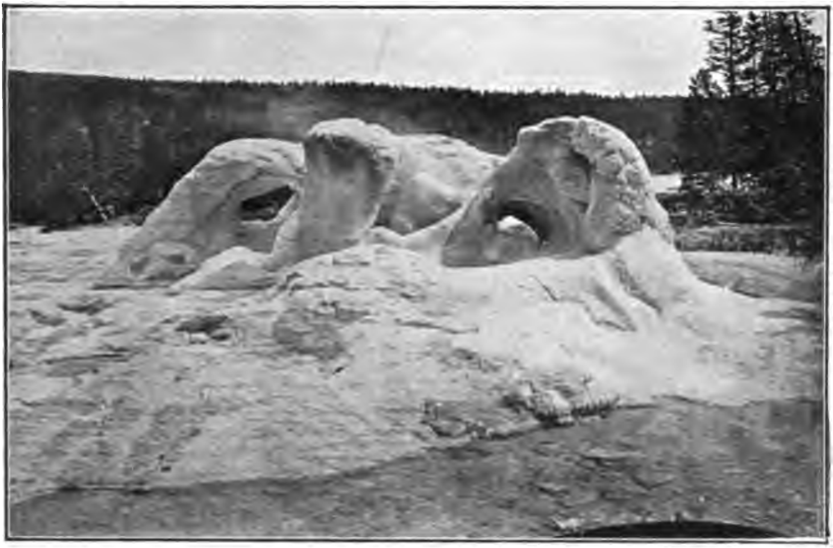

Fig. 216.—Cone (or crater) of Grotto Geyser, Yellowstone Park. (Detroit Photo. Co.)

about themselves, the cone being of material deposited from solution. In the Yellowstone Park the precipitation of the matter in solution (chiefly silica) is partly due to cooling and partly to the algæ which abound even in the boiling water, and the brilliant colors of the deposits about the springs are attributable to these plants. When the water from any geyser or hot spring ceases to flow the plants die and the colors disappear. The details of the surface of the deposits about geysers and hot springs are often complicated, and frequently very beautiful (Fig. 218).

The heating of geyser and hot-spring water must cool the lava or

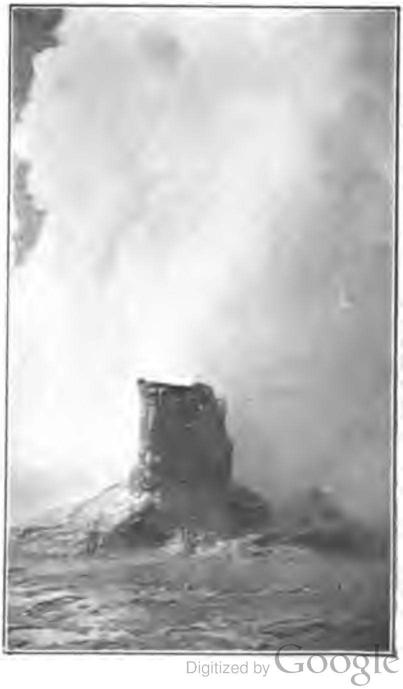

Fig. 217.—Cone of Giant Geyser, Yellowstone Park. (Wineman.)

other source of heat below. As this takes place, the time between eruptions becomes longer and longer. In the course of time, therefore, the geyser must cease to be eruptive, and when this change is brought about the geyser becomes a hot spring. Within historic times several geysers have ceased to erupt and new ones have been developed. In the Yellowstone Park, where there are said to be something like 3000 vents of all sorts, hot springs which are not eruptive greatly outnumber the geysers. From many of the vents but little steam issues, and from some, little else

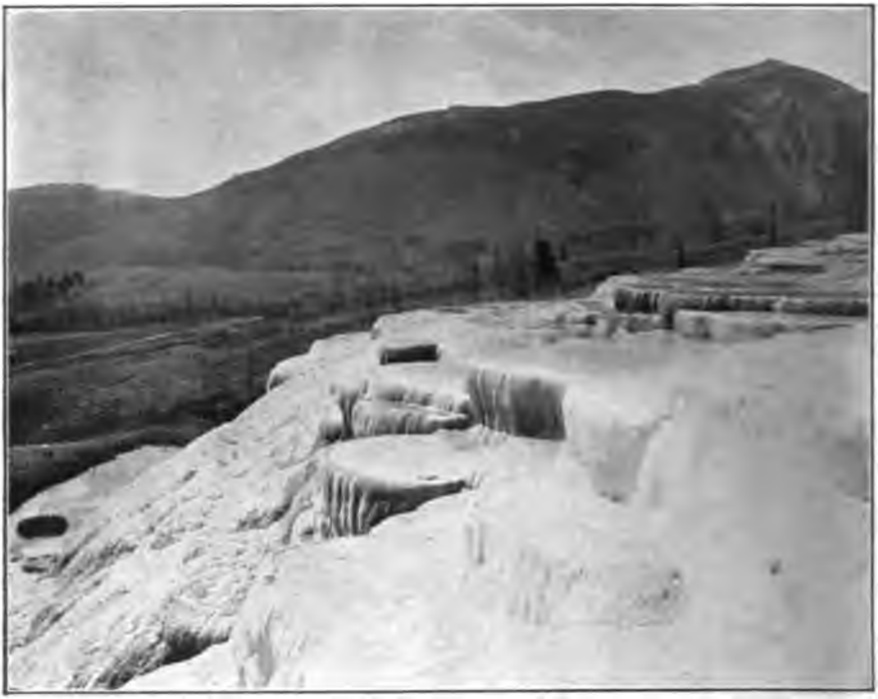

Fig. 218.—Hot springs deposits. Terraces of Mammoth Hot Springs, Yellowstone Park.

A few geysers have somewhat definite periods of eruption. Of such "Old Faithful" is the type; but even this geyser, which formerly erupted at regular intervals of about an hour, is losing the reputation on which its name is based. Not only is its period of eruption lengthening, but it is becoming irregular, and the irregularity appears to be increasing. In the short time during which this geyser has been under observation its period has changed from a regular one of sixty minutes, or a little less, to an irregular one of seventy to ninety minutes. In the case of some geysers years elapse between eruptions, and in some the date of the last eruption is so distant that it is uncertain whether the vent should be looked upon as a geyser or merely a hot spring.

In the Yellowstone Park[1] the geysers are mainly in the bottoms of valleys (Fig. 219), but the deposits characteristic of geysers are found in not a few places well above the present bottoms. These deposits record the fact that in earlier times the geysers were at higher levels than now. It is probable that they have been, at all stages in their history, near the bottoms of the valleys, and that, as the valleys have been deepened the ground-water has found lower and lower points of issue. In this respect the geysers have probably had the same history as other springs.

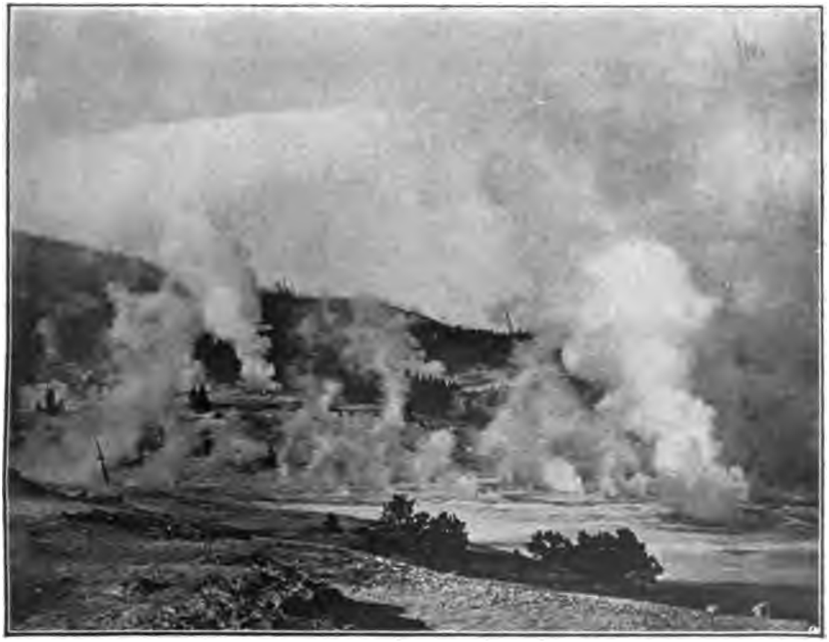

FIG 219.—Hot springs and geysers. Norris Geyser basin, Yellowstone Park.

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[1] Geikie. Geological Sketches, pp. 206–38. Hayden. Amer. Jour. Sci., Vol. III, 1872, pp. 105–15 and 161–76.

Unless new intrusions of lava occur, or unless heat is otherwise renewed at the proper points, it is probable that all existing geysers will become extinct within a time which is, geologically, short. New geyser regions may, however, develop as old ones disappear.

Artesian wells.—Originally the terms *artesian* wells and *flowing* wells were synonymous; but at the present time any notably deep well is called *artesian*, especially if it descends to considerable depths below the mantle rock. The artesian well which does not flow, does not differ from common wells in principle; but being deeper, the water which it affords is often more thoroughly filtered and frequently more highly mineralized than that of other wells. The flowing well is really a gushing spring, the opening of which was made by man.

Flowing wells[1] depend upon certain relations of rock structure, water supply, and elevation. Generally speaking a flowing well is possible in any place underlain by any considerable bed of porous rock, if such rock outcrops at a sufficiently higher level in a region of adequate rainfall, and is covered by a layer or bed of impervious, or relatively impervious rock. This statement involves four conditions, all of which are illustrated by Fig. 199, where *a* is the bed of porous rock. It is not necessary that the beds of rock form a structural basin, nor is it usually necessary to take account of the character of the rock beneath the porous bed which contains the water.

The bed of porous rock is the "reservoir" of the flowing well. Formations of sand or sandstone, and of gravel or conglomerate, most commonly serve as the reservoirs. In order that it may contain abundant water it must have some thickness, and its outcropping edge must be so situated that the water may enter freely and be replenished, chiefly by rain, as the water flows out at the well.

A relatively impervious layer of rock above the reservoir (*b*, Fig. 199) is most important; otherwise the water in the reservoir will leak out, and there will be little or no "head" at the well site. Thus if the rock overlying stratum *a* (Fig. 199) were badly broken, the fractures extending up to the surface, the conditions would be unfavorable for flowing wells. Under such conditions, wells in the positions of those shown in Fig. 199 might get abundant water, but they would not be likely to flow. If the stratum next below the reservoir is not impervious, some lower one probably is. No layer of rock is more impervious than one which is full of water, and the substructure of any bed which might serve as a reservoir is usually full of water, even if the rock be porous.

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[1] Chamberlin, Geol. of Wis., Vol. I, pp. 689–97, and Fifth Ann. Rept., U. S. Geol. Surv., pp. 131–73. The former a brief, and the latter an elaborate, exposition of the principles involved.

If the outcrop of the reservoir be notably above the site of the well, and if it be kept full by frequent rains, the "head" will be strong, though the water at the well will not rise to the level of the outcrop of the reservoir. Experience has shown that an allowance of about one foot per mile of subterranean flow should be made. Thus if the site of the well be 100 miles from the outcrop of the water-bearing stratum, and 200 feet below it, the water will rise something like 100 feet about the surface at the well. This rule is, however, not applicable everywhere. The failure of the water to rise to the level of its head is due to the adhesion and the friction of flow through the rock. The more porous the rock the less the reduction of head by friction. The height of the flow is also influenced by the number of wells drawing on the same reservoir, on the degree of imperviousness of the confining bed above, etc.

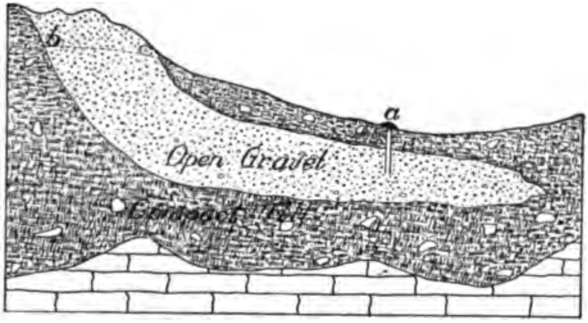

Fig. 220.—Figure illustrating the principle of artesian wells in drift.

Flowing wells, often relatively shallow, are frequently obtained from unconsolidated drift. Some such relations as suggested by Fig. 220 would afford the conditions for flowing wells in such a formation.

CHAPTER V.

THE WORK OF SNOW AND ICE.

A PART of the atmospheric precipitation falls as snow, and this, like the rain, does its appropriate work in degrading the land. Over the larger part of the land surface the snow of the winter does not endure through the succeeding summer, and when it melts it follows the same course as the precipitation which falls as rain; but in cold regions where the fall of snow is heavy some of it remains unmelted and constitutes perennial snow-fields.

SNOW- AND ICE-FIELDS.

Snow-fields.—Mountain heights and polar lands are the most common habitats of snow-fields, though they are not confined to these situations. In North America there are numerous small snow-fields in the western mountains, from Mexico on the south to Alaska on the north, their number and size increasing in the latter direction. In the United States there are few snow-fields south of the parallel of 36° 30′, and most of the many hundreds north of that latitude (excluding Alaska) are small (Pl. XVIII, Fig. 1, Washington, lat. 48° 5′, long. 121° 5′; Fig. 2, lat. 41° 25′, long. 122° 12′. From Glacier Peak and Shasta Special Quadrangles, U. S. Geol. Surv.). Farther north, especially in Alaska, the snow-fields of the western mountains attain much greater size. In Europe snow-fields comparable to those of the northwestern part of the United States and British Columbia occur in the Alps (Fig. 221), the Pyrenees, the Caucasus, and the Scandinavian mountains. In Asia snow-fields occur in the Himalayas and in many of the high mountains farther north, from Turkestan on the southwest nearly to the coast on the northeast. In South America there are snow-fields of small size even in equatorial latitudes, and farther south in the Chilean Andes there are some of considerable size. Small snow-fields occur on the highest peaks of tropical Africa, and in the mountains of New Zealand. For reasons which will appear later, much of every considerable snow-field is really ice.

In addition to these limited fields of snow in mountain regions, there are fields of much greater extent covering wide expanses of plain and plateau in the polar regions. The greater part of the island of Greenland is covered with a single field of ice and snow, the size of which is variously estimated at 300,000 to 400,000 square miles (Fig. 222)—an area 400 to 600 times as large as the snow-and-ice-covered area of Switzerland. Numerous islands to the west of North Greenland are also partly covered with snow, the areas of the snow-fields far exceeding those of most mountain regions. In Antarctica there is believed to be a still larger field, the largest of the earth. Its area is not even approximately known, but such data as are at hand indicate that it may have an extent of 3,000,000 or 4,000,000 square miles.

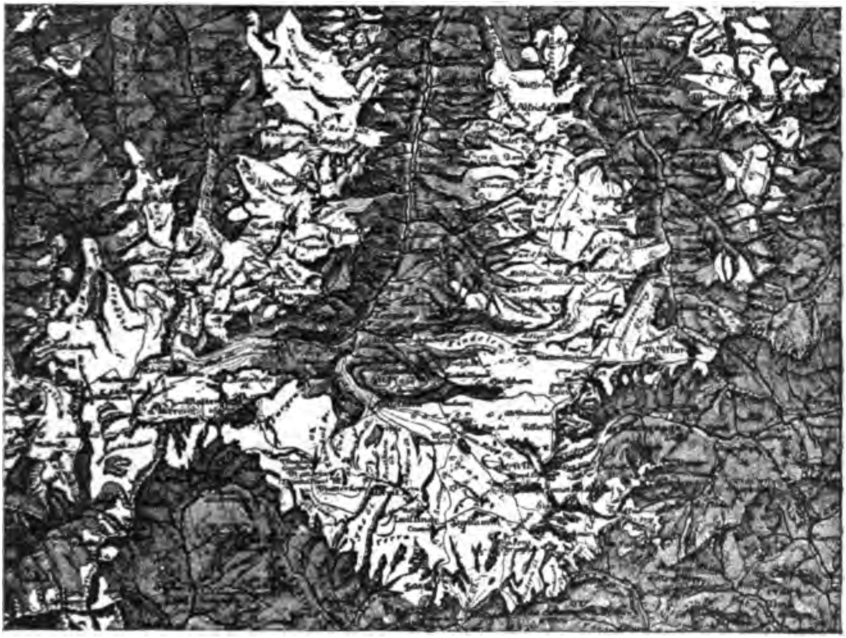

Fig. 221.—An alpine snow-field.

The only condition necessary for a snow-field is an excess of snowfall over snow waste. The lower edge of a snow-field, the *snow-line,* is dependent chiefly on temperature and snowfall. In general it does not depart much from the summer isotherm of 32°, though it may be well above this isotherm where the snowfall is light. That the snow-line is not a function of temperature only is shown by its position in various places. In the equatorial portion of the Andes, for example, the snow-line has an altitude of about 16,000 feet on the east side of the mountains, where the precipitation is heavier, and of about 18,500 feet on the west side, where it is lighter. For the same reason the snow-line in the Himalayas is 3000 or 4000 feet lower on the *south* side than on the north.

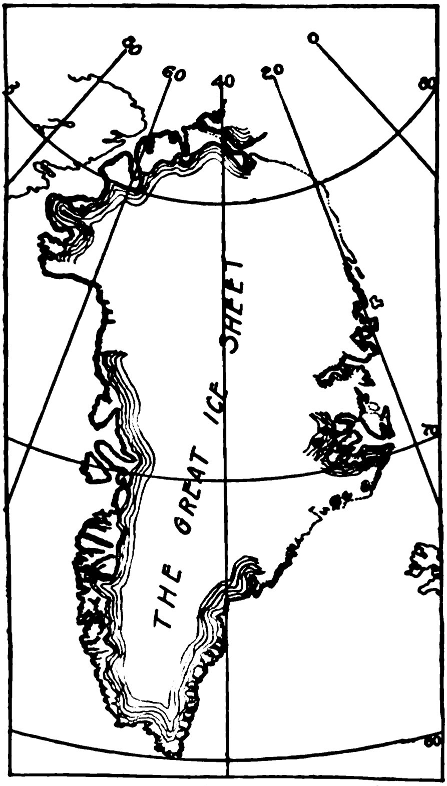

FIG. 222.—Map of Greenland. The borders only are free from ice. (Stieler.)

While in equatorial regions the snow-line has an altitude of 15,000 to 18,000 feet, it approaches or even reaches sea-level in the latitude of Antarctica and North Greenland. In intermediate latitudes it has an intermediate position.

While temperature and snowfall are the most important factors controlling the position of the snow-line, both humidity and the movements of the air are of some importance, since both affect the rate of evaporation of snow and ice.

The passage of snow into névé and ice.—The snow does not lie on the surface long before it undergoes obvious change. The light flakes soon begin to be transformed into granules, and the snow becomes "coarse-grained." The granular character, so pronounced in the snow of the last banks which remain in the spring in temperate latitudes, is even more distinct in perennial snow-fields, either at the surface or just beneath it. This granular snow is called *névé*, or *firn*. Still deeper beneath the surface, where the thickness of the snow is great, the névé becomes more compact and finally coherent, and grades into porous ice. This gradation is accomplished at no great depth, though the thicknesses of snow and névé are by no means constant.

Structure of the ice.—Ice formed beneath a snow-field is in some sense stratified. It is made up of successive falls of snow which tend to retain the form of layers. This follows from two or three conditions. The snow of one season, or of one period of precipitation, may have been considerably changed before the succeeding fall of snow. So also the surface of the snow-field at the end of the melting season is often covered with a visible amount of earthy matter, some of which was blown up and dropped on the surface during the melting season, and some of which was concentrated in that position by the melting of the snow in which it was originally imbedded. The amount of earthy matter is often sufficient to define snows of successive years, or perhaps of minor periods of precipitation, and makes distinct the stratification which would otherwise remain obscure. The snowfall of successive years has been estimated by this means[1] where the snow is exposed in crevices in the snow-field.

In addition to its rude stratification, the ice of the deeper portions often acquires a stratiform structure which may perhaps best be called *foliation* to distinguish it from the *stratification* which arises from deposition. The foliation appears to result mainly from the shearing of one part over another in the course of the movements to which the ice is subjected, as will be illustrated presently.

Texture.—The ice derived from the snow is formed of interlocking crystalline grains. The crystalline character is present from the beginning, for it is assumed by the snowflakes when they form, and the subsequent changes seem only to modify the original crystals by building up some and destroying others. By the time the snow is converted into névé, the granules have become coarse, and wherever the ice derived from the névé has been examined, the granular crystalline texture is present. The individual crystals in the ice are usually larger than those of the névé, and more closely grown together. In the fresh unexposed ice the crystals are so intimately interlocked that they are not readily seen except under a polarizing microscope, but when the ice has been honeycombed by partial melting, the granules become partially separated and may be easily seen. Fig. 223 shows quantities of them which have been washed down from the surface, and disposed as cones at its base. While a given mass of snow in a great snow- and ice-field cannot be followed consecutively through its whole history, yet since (1) the granular texture is pronounced in the névé stage where the granules show evidences of growth, and since (2) the same texture is also pronounced in the last stages of the ice when it is undergoing dissolution, as well as at all observed intermediate stages, and since (3) the crystals are, on the average, larger in proportion to the lateness of the stage of their history, while (4) experiment has shown that granules grow under the conditions which exist in snow-fields, and (5) that they persist under very considerable pressure, it is legitimate to assume that a granular crystalline condition persists throughout all stages, and is a feature of progressive growth.

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[1] Russell. Nat'l Geog. Mag., Vol. III, pp. 127 and 181.

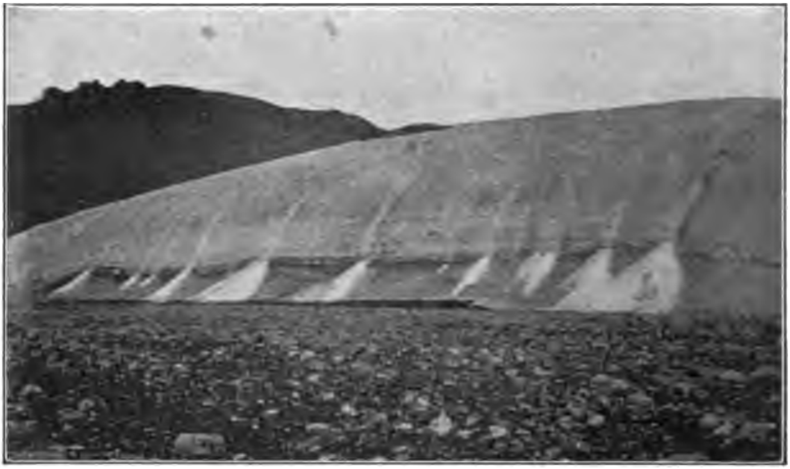
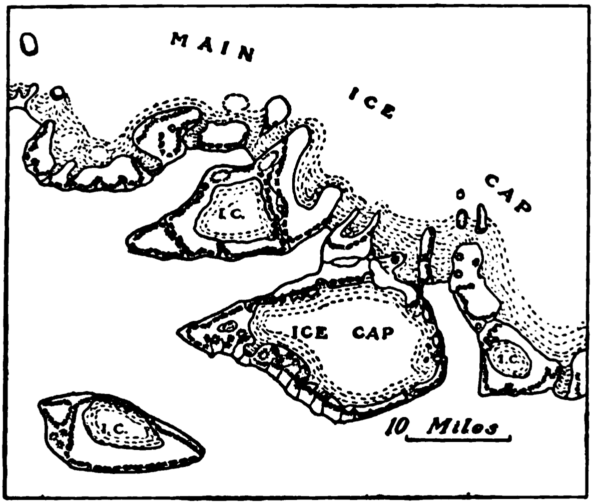
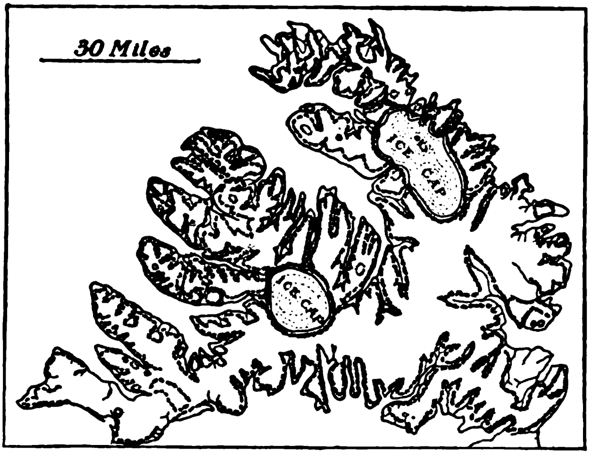


Fig. 223.—Figure showing cones of granules of ice which have been washed down the front of the glacier by streamlets, and accumulated after the manner of talus or alluvial cones. North Greenland.

Inauguration of movement.—Eventually the increase in depth of snow and ice in a snow-field gives rise to motion. The exact nature of the motion has not yet been demonstrated to the satisfaction of all investigators. Brittle and resistant as ice seems, it exhibits, under proper conditions, some of the outward characteristics of a plastic substance. Thus it may be made to change its form, and may even be moulded into almost any desired shape if carefully subjected to sufficient pressure, steadily applied through long intervals of time.[1] These



ɪɢ. 224.—Ice-caps of small size. The figure also shows some valley glaciers extending out from the main ice-sheet and from the local ice-caps. A portion of the North Greenland coast, north of Inglefield Gulf. Lat. about 78°. (Peary.)



ɪɢ. 225.—Small ice-caps in the northwestern part of Iceland. (Thoraddsen's geological map of Iceland.)

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[1] For an account of experiments illustrating the mobility of ice see Aitkin, Am. ur. Sci., Vols. V, p. 303, and XXXIV, p. 149, and Nature, Vol. XXXIX, p. 203.

changes may be brought about without visible fracture, and have been thought to point to a viscous condition of the ice. There is much reason, however, as will be seen later, to question this interpretation of the ultimate nature of the movement. Whatever this may be, the *mass result* of the movement in a field of ice is comparable, in a superficial way at least, to that which would be brought about if the ice were capable of moving like a viscous liquid, the motion taking place with extreme slowness. This slow motion of ice in an ice-field is *glacier motion*, and ice thus moving is *glacier ice.*

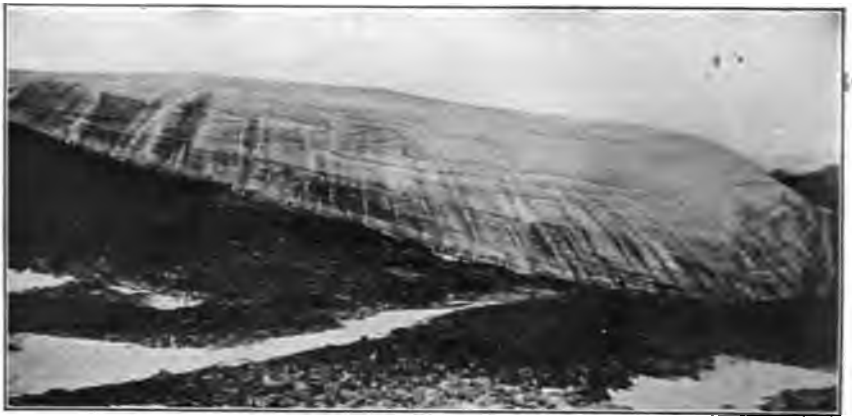

Fig. 226.—A glacial lobe, midway between an ice-cap and a valley glacier. A protrusion from a local ice-cap east of Cape York, Greenland.

If both the surface on which the ice-sheet develops and its surroundings be essentially plane, as may happen in high latitudes, and if the snow- and ice-field be symmetrical in shape, the outward movement will be approximately equal in all directions, and the area covered by the spreading ice-field will remain more or less circular. If the ice-field rests on a steeply inclined surface, like a mountain slope, the movement becomes one-sided in conformity to the slope. If the surface, otherwise plane, be affected by valleys parallel to the direction of movement, the ice in the valleys will be deeper than that on the divides between them, and its movement stronger. In the valleys, therefore, the ice will advance farther than elsewhere before being melted, and the outline of the ice will become lobate, the lobes occupying the depressions. These general relations are shown in Figs. 224 and 225. If the depressions be wide and shallow, the lobes will be broad and short

(Fig. 226); if the depressions be narrow and deep, the lobes will be relatively narrow and long. If the snow and ice rest on a surface consisting chiefly of steep valleys and sharp ridges, as is common on mountain slopes, the snow and ice are chiefly gathered in the valleys, and take a linear form.

TYPES OF GLACIERS.

These different forms give rise to different terms. The ice which spreads with some approach to equality in all directions from a center is a glacier, is indeed the type of the greatest glaciers, but is commonly called an *ice-cap*. The same name is applied to any glacier in which there is movement in all directions from the center, even though its shape departs widely from a circle. The glacier covering the larger part of Greenland (Fig. 222) is a good example of a large ice-cap, and the glaciers on some of the flat-topped peninsular promontories of the same island are good examples of small ones (Fig. 224). If ice-caps cover a large part of a continent, as some of those of the past have done, they are often called *continental glaciers*.

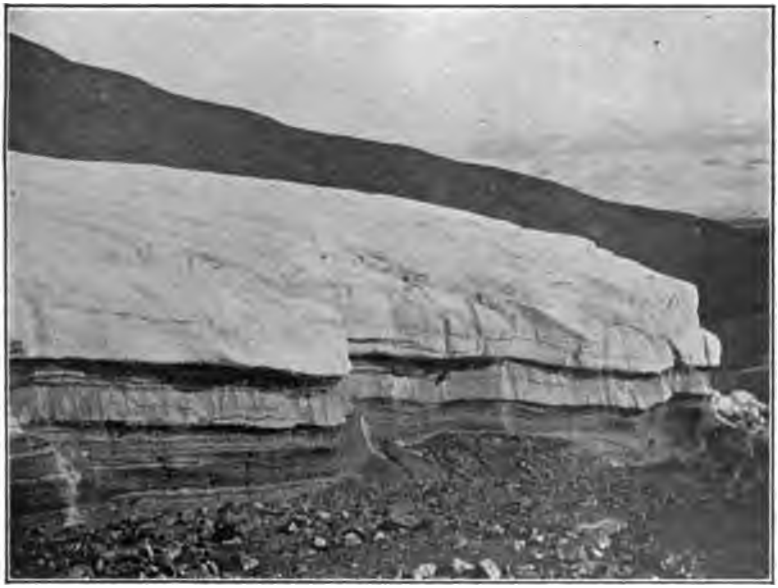

FIG. 227.—Characteristic end of a North Greenland (Bryant) glacier.

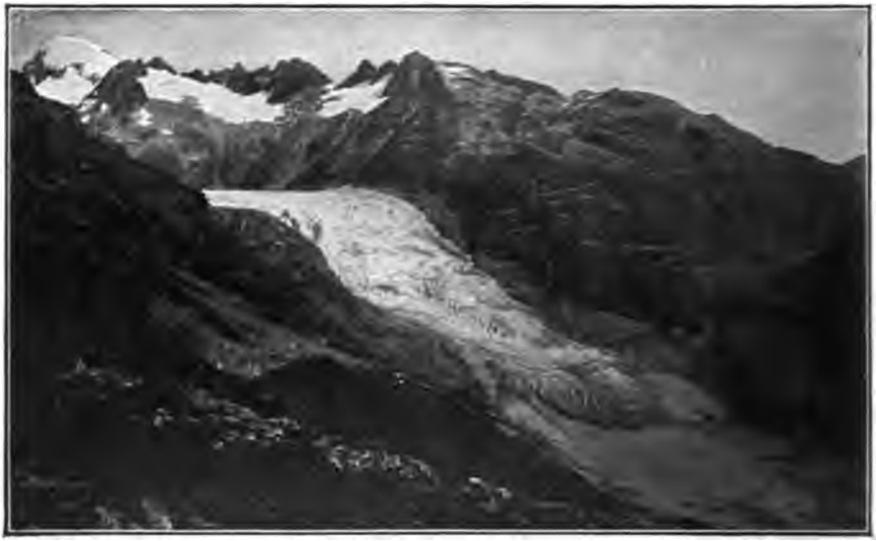

Fig. 228.—The Rhône glacier. (Reid.)

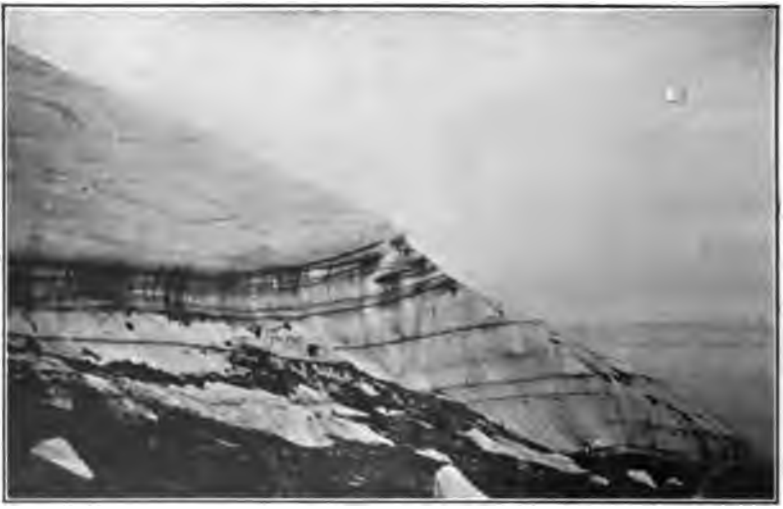

Fig. 229.—Characteristic end of a North Greenland glacier. North side of Herbert Island, Inglefield Gulf.

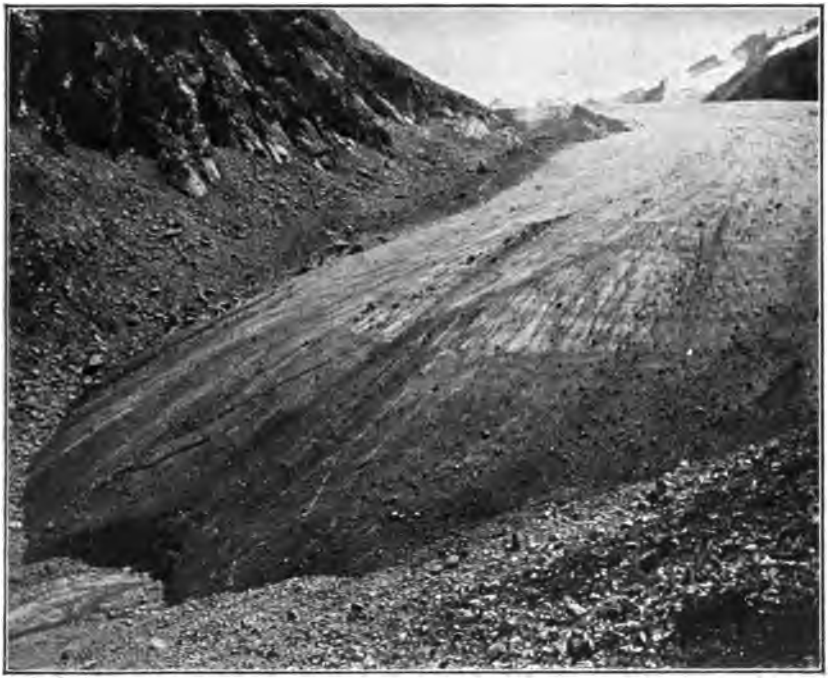

FIG. 230.—The end of an alpine (Forno, Switzerland) glacier. (Reid.)

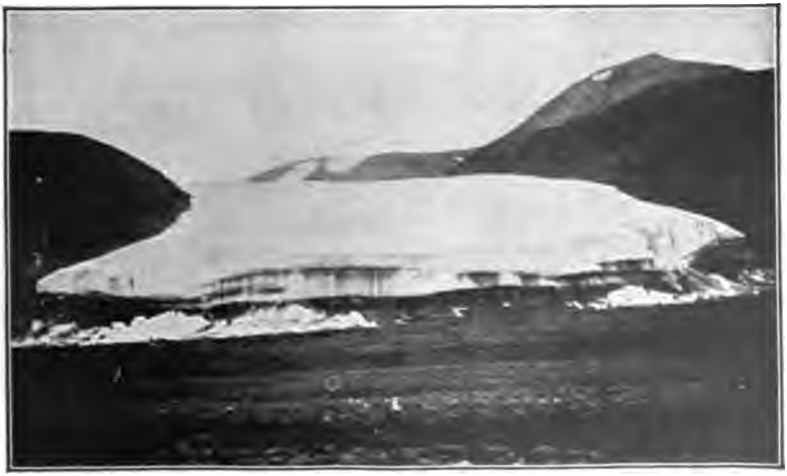

FIG. 231.—Deploying end of a North Greenland glacier.

Where ice-caps are developed on plateaus whose borders are trenched by valleys, ice-tongues from the edge of the ice-cap often extend down into the valleys and give rise to one type of *valley glacier* (Figs. 224 and 227). A second and more familiar type of valley glacier occupies mountain valleys, and is the offspring of mountain snow-fields (Fig. 228). The former are confined chiefly to high latitudes, and are distinguished as *polar* or *high-latitude* glaciers (Figs. 227 and 229); the latter are known as *alpine* glaciers (Figs. 228 and 230). The end and side slopes of polar glaciers are, as a rule, much steeper than those of alpine glaciers. When a valley glacier descends through its valley to the plain beyond, its end deploys, forming a fan (Fig. 231). The deploying ends of adjacent glaciers sometimes merge, and the resulting body of ice constitutes a *piedmont glacier* (Fig. 232). At the present time, piedmont glaciers are confined to high latitudes. In some cases the snow-field that gives rise to a glacier is restricted to a relatively small depression in the side of a mountain, or in the escarpment of a plateau. In such cases the snow-field and glacier are hardly distinguish-

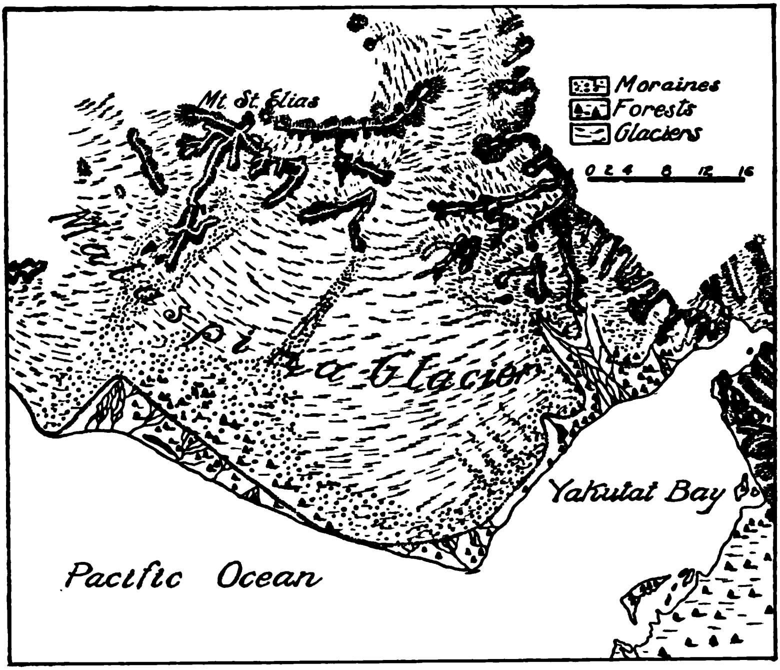

FIG. 232.—The Malaspina glacier, Alaska; the best known example of a piedmont glacier. (Russell.)

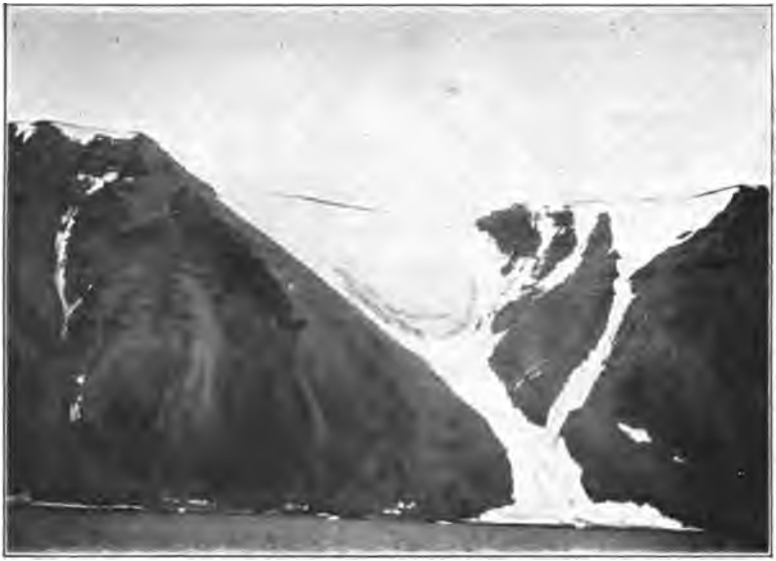

FIG. 233.—A cliff glacier. North Greenland type. North side of Herbert Island, Inglefield Gulf. The lower half of the white area is snow, and snow talus. So also are the white patches to the right. The height of the cliff is perhaps 2000 feet. The water in the foreground is the sea.

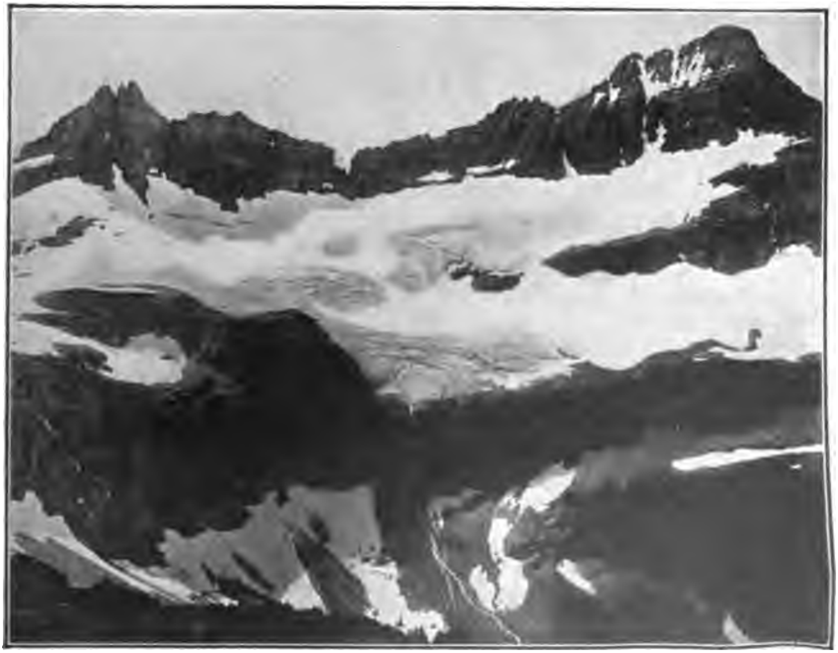

FIG. 234.—Chaney glacier; a cliff glacier of the Montana type. (Shepard.)

able, and the latter descends but little below the snow-line. In many cases it does not even enter the narrow valley which leads out from the depression occupied by the snow-field. Such a glacier is nestled in the face of a cliff, and may therefore be called a *cliff glacier*[1] (Figs. 233 and 234). The snow-field of a cliff glacier is sometimes no more than a great snowdrift, accumulated through successive years. Cliff glaciers are often as wide as long, and are always small, and between them and valley glaciers there are all gradations (Fig. 235). Occasionally the end of a valley glacier, or the edge of an ice-sheet reaches a precipitous cliff, and the end or edge of the ice breaks off and accumulates like talus below. The ice fragments may then again become a coherent mass by regelation, and the whole may resume motion. Such a glacier is called a *reconstructed glacier*. The precipitous cliffs of the Greenland coast furnish illustrations.

Of the foregoing types of glaciers, the ice-caps far exceed all others both in size and importance, while the valley glaciers out-rank, in the same respects, the other types; but since the valley glaciers are the most familiar type, the general phenomena of glaciers will be discussed with primary reference to them.

THE GENERAL PHENOMENA OF GLACIERS.[2]

Dimensions.—Glaciers which occupy valleys leading down from snow-fields sometimes reach the upper parts of the valleys only, sometimes extend through them, and sometimes push out on the plain beyond. In length they range from a fraction of a mile to many miles, and though their width is usually much less than their length, the reverse is sometimes the case (Figs. 233, 234, and 235). Their thickness is usually measured by hundreds of feet rather than by denominations

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[1] Jour. of Geol., Vol. III, p. 888.

[2] The following list includes many of the more available articles and treatises on existing glaciers; others are referred to in the following pages.

Alaskan glaciers: Reid, (1) Nat. Geog. Mag., Vol. IV, pp. 19–55; (2) Sixteenth Ann. Rept., U. S. Geol. Surv., Part I, pp. 421–461. Russell, (1) Nat. Geog. Mag., Vol. III, pp. 176–188; (2) Jour. of Geol., Vol. I, pp. 219–245.

Glaciers in the United States: Russell, (1) Fifth Ann. Rept., U. S. Geol. Surv., pp. 309–355; (2) Eighteenth Ann. Rept., U. S. Geol. Surv., Part II, pp. 379–409; (3) Glaciers of North America.

Greenland glaciers: Chamberlin, Jour. of Geol., Vol. II, pp. 768–788; Vol. III, pp. 61–69, 198–218, 469–480, 565–582, 668–681, and 833–843; Vol. IV, pp. 582–592. Salisbury, Jour. of Geol., Vol. III, pp. 875–902, and Vol. IV, pp. 769–810.

Glaciers in general: Shaler and Davis, Illustrations of the Earth's Surface; Forbes, Norway and its Glaciers, and Theory of Glaciers; Heim, Handbuch der Gletscherkunde.

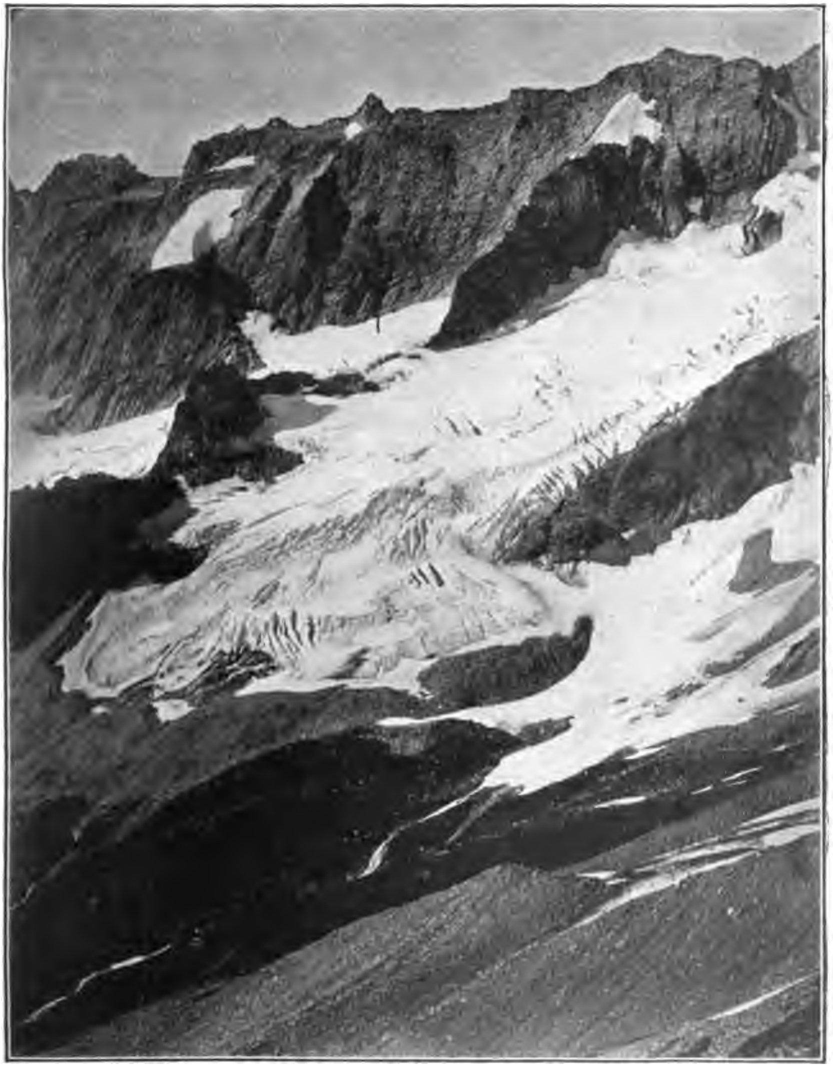

FIG. 235.—A glacier in the Cascades near Cascade Pass, Wash. A glacier intermediate between a cliff glacier and a valley glacier. (Willis, U. S. Geol. Surv.)

of other orders, but the variation is great, and exact measurements are almost wholly wanting. The minimum thickness is that necessary to cause movement, and this varies with the slope, the temperature, and other conditions. There is also much variation in the thickness in different parts of a glacier. As a rule, it is thinnest in its terminal portion, and thickest at some point intermediate between this and its source, but nearer the latter than the former. Cliff and reconstructed glaciers are comparable in size to the smaller valley glaciers. Piedmont glaciers may attain greater size.

An ice-cap is theoretically thickest at its center and thins away to its borders, but its actual dimensions are influenced by the topography on which it is developed. The Greenland ice-cap is known to rise about 9000 feet above the sea, and it probably reaches considerably higher than this in the unexplored center of its broad dome. The height of the land surface beneath is unknown, but it is unlikely that it averages half this amount, and hence the ice is probably 5000 feet or more thick in the center. There is reason to think that it is much thicker in Antarctica.

Limits.—The ice of a glacier is always moving forward (neglecting temporary halts), but the *end* of a glacier may be retreating, advancing or remaining stationary, according as the rate of wastage is greater, less, or just equal to the forward movement of the ice. The position of the lower end of the glacier is therefore determined by the ratio of movement to wastage. Its upper end is generally ill-defined. In a superficial sense, it is the point where the ice emerges from the snow-field; but the lower limit of the snow-field is often ill-defined, and in any case is not the true upper limit of the glacier, since there must be movement from the granular mass of ice beneath the snow to make up for the waste below, and the moving ice beneath the snow-field which feeds the tongue of ice in the valley is just as really a part of the glacier as the more consolidated portion in the valley below. If a definite upper limit for an alpine glacier is to be named, it should probably be the *Bergschrund*, a gaping crevasse, or series of crevasses which sometimes open near the precipitous slope of the peak or cliff where the snow-field lies. The *Bergschrund* is formed by the moving of the lower part of the snow-field away from the portion above.

The lower end of a glacier is usually free from snow and névé in summer, but, traced toward its source, it first becomes covered with névé, then with snow, and finally merges into the snow-field without having ceased to be a glacier. The term glacier is, however, commonly used to mean merely the more solid portion outside (below) the névé.

Movement.—The fact of glacier movement is established in various ways, the most obvious being by the advance of its lower end. Such advance is too slow to be seen from day to day, and is only detected when the lower end of the glacier overrides or overturns objects in front of it, or moves out over ground previously unoccupied. But even when the end of a glacier is not advancing, the movement of the ice may be established by means of stakes or other marks set on the surface. If the positions of these marks relative to fixed points on the sides of the valley be determined, they are found after a time to have moved down the valley. Rows of stakes or lines of stones set across a glacier in the upper, middle and lower portions have revealed many facts concerning the movement of the ice.

Generally speaking, the middle of a valley glacier moves more rapidly than its sides (Fig. 236), but in some cases, especially in large glaciers, there are found to be two or more main lines of movement, with belts of lesser movement between. The top of a glacier moves, on the whole, more rapidly than the bottom, though the observations made do not show that the rate of movement diminishes regularly downward, and it probably does not so diminish in many cases. In Switzerland, where the glaciers have been studied more carefully than elsewhere, the determined rates of movement range from one or two inches to four feet or more per day. Some of the larger glaciers in other regions move more rapidly, but it does not follow that large glaciers always move faster than small ones. The Muir glacier of Alaska has been found to move seven feet or more per day,[1] and some of the glaciers of Greenland have been found to move, in the summer time, 50, 60, or even more feet per day. A single estimate as high as 100 feet per day has been made; but these high rates have been observed only where the ice of a large inland area crowds down into a comparatively narrow fjord, and debouches into the sea, and then only in the summer. In the case of the glacier with the highest recorded rate of summer movement, 100 feet per day, the advance was only 34 feet at about the same place in April.

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[1] Reid. Natl. Geog. Mag., Vol. IV, p. 44.

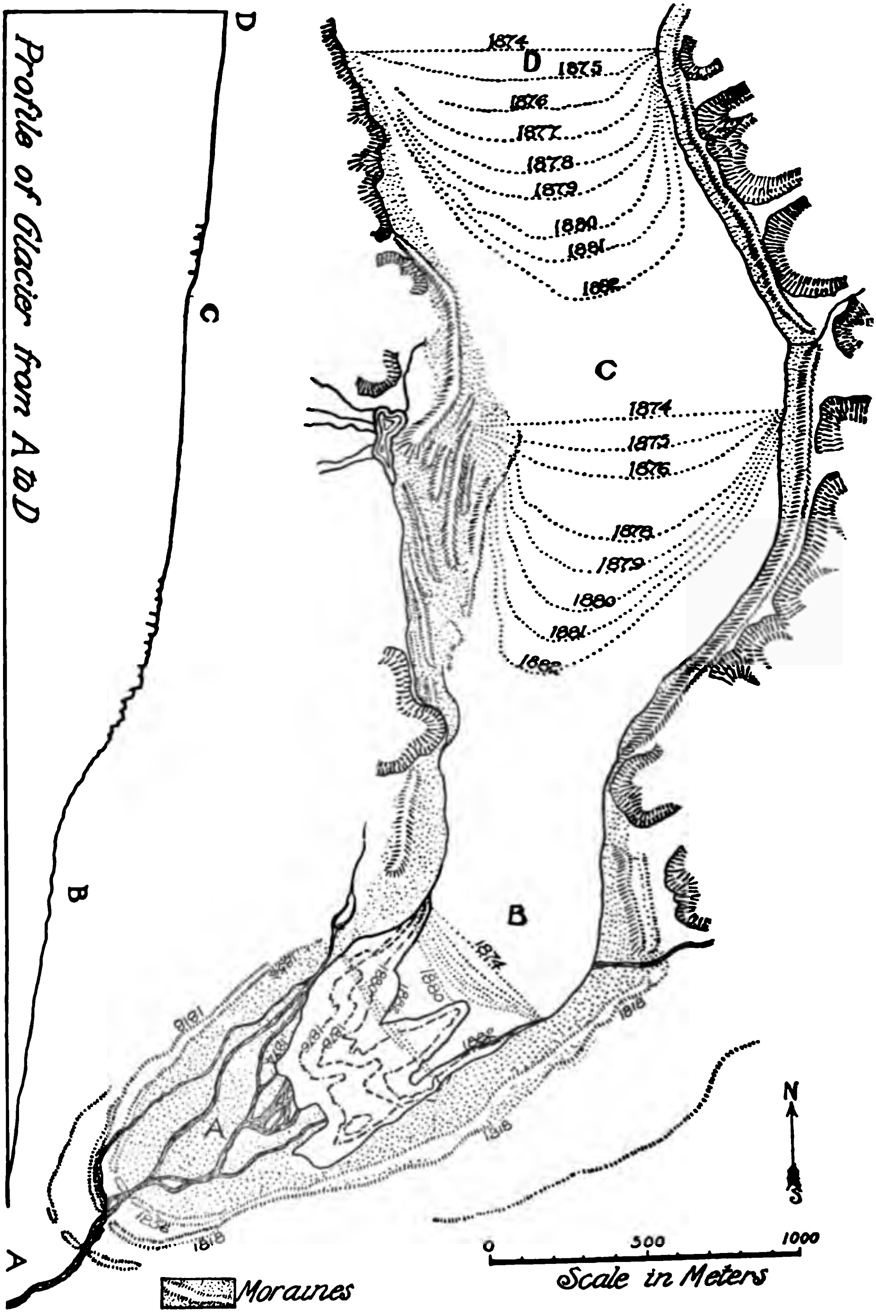

Fig. 236.—Diagram to show the rate of movement of the Rhone Glacier at various points in its course at centre and sides. It also shows the fluctuations in the positions of the end of the glacier between 1874 and 1882, and the profile of the ice. (Heim.) Page 260.

The *average* movement of the border of the inland ice of Greenland is very small. Rink says that "between 62° and 68° 30′, the edge of the inland ice is almost stationary for a remarkably long distance."[1] The observations of the authors between 77° and 78° were of like import. Probably the average movement of the border of the Greenland ice-cap is less than one foot a week.

Conditions affecting rate of movement.—The rate of glacier movement appears to depend on (1) the depth of the moving ice; (2) the slope of the surface over which it moves; (3) the slope of the upper surface of the ice; (4) the topography of the bed over which it passes; (5) the temperature; and (6) the amount of water which falls upon it or is carried to it by the drainage of its surroundings, in addition to that produced by the melting of the glacier itself. Great thickness, a steep slope, much water, smoothness of bed, and a high (for ice) temperature favor rapid movement. Since some of these conditions, notably temperature and amount of water, vary with the season, the rate of movement for any given glacier is not constant throughout the year. Other conditions, especially the first of those mentioned above, vary through longer periods of time, and occasion periodic variations in the rate of movement.

Since the volume of ice concerned influences the rate of movement, anything which changes the volume affects the rate. An excess of snowfall with favorable conditions for its preservation for a period of years, increases the volume of ice, and tends to accelerate its movement. A deficiency in snowfall, or in its preservation, as from high average temperature or from aridity, diminishes the quantity of ice, and so retards the movement. An acceleration of velocity causes the ice to move down the valley farther before being melted, that is, causes the end of the glacier to advance, while a decrease of velocity produces the opposite effect. As a matter of fact, the lower ends of glaciers advance for a period of years and then retreat, to advance again at a later time.[2] Observation has shown that the periods of advance follow a succession of years when the snowfall has been heavy and the temperature low, while the periods of retreat follow a succession of years when the snowfall has been light and the temperature above the average. The periods of advance and retreat lag behind the periods of heavy and light snowfall respectively, by some years, and a long glacier responds less promptly than a short one. Present knowledge seems to point to a period of 35 to 40 years as the time within which a cycle of fluctuation, that is, an advance and a retreat, takes place.

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[1] Rink's Greenland.

[2] Reid. Variations of Glaciers. Jour. of Geol., Vols. III, p. 278; V, p. 378; VI, p. 473; VII, p. 217; VIII, p. 154; IX, p. 250, and X, p. 313.

A declining upper surface is essential to glacier motion. There are short stretches where this is not the case, and indeed there are particular places where the upper surface slopes backward.[1] This may occur where the ice is pushed up over a swell in its bed, or is crowded up against any considerable obstacle; but such cases are no more than local exceptions, and do not militate against the truth of the general statement that the upper surface of a glacier declines in the direction of motion. A declining *lower* surface is less necessary. In the case of valley glaciers, the bed does, as a rule, decline in the direction of motion, but that there are local exceptions is shown by the deep basins in rock which such glaciers often leave behind them when they retreat. In the great continental glaciers of recent geologic times, the ice frequently moved up slopes for scores, and even hundreds of miles; but in all such cases, the upper surface must have declined in the direction of movement. With a given thickness of ice, the greater the decline of its lower surface in the direction of motion, the more rapid its progress. A rough bed, or a crooked course retards the motion of a glacier, while a smooth bottom and a straight course facilitate it.

Slope, roughness of bed, and volume affect the movement of glaciers somewhat as they affect the movement of rivers. The temperature of the water, on the other hand, has little effect on the flow of a river so long as it remains unfrozen; but the effect of temperature on the motion of ice is most important. In many cases, indeed, the temperature, together with the water that is incidental to it, seems to be the chief factor in determining the rate of movement. The way in which its effects are felt will be discussed later.

Likenesses and unlikenesses of glaciers and rivers.—Many of the characteristics of a valley glacier may be understood from the study of the accompanying figure (Fig. 237) of the White (Alaska) glacier. From this figure it will be seen that the glacier is an elongate river-like body, following the curves of the valley in stream-like fashion. It has its origin in the snows collected on the mountain heights seen in the distance, and it works its way down the valley in a manner which, in the aggregate, is similar to the movement of a stiff liquid. The likeness to a river extends to many details. Not only does the center move faster than the sides, and the upper part faster than the bottom, as in the case of streams, but the movement is more rapid in constricted portions of the valley and slower in the broader parts. These and other likenesses, some of which are apparent rather than real, have given origin to the view that glacier ice moves like a stiff viscous liquid.

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[1] For example, in the Middle Blase Dale glacier, Island of Disco, Jour. of Geol., Vol. II, p. 784, and in the Bowdoin glacier (Fig. 242).

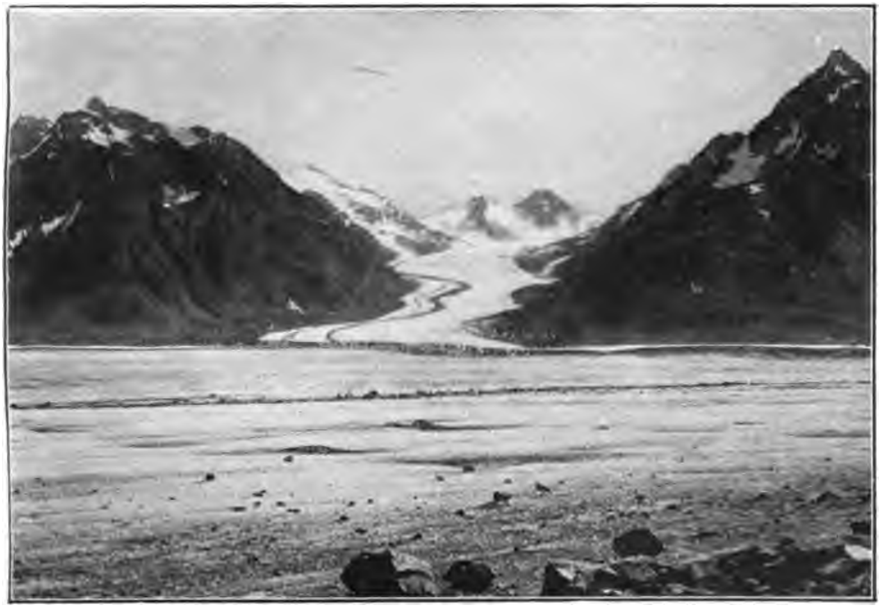


FIG. 237.—White glacier (central background) joining a larger glacier (foreground), Alaska. (Reid.)

But while the points of likeness between glaciers and rivers are several, their differences are at least equally numerous and significant. The trains of débris on the surface (the dark bands in the illustration), like the central currents of streams, pass nearer the projecting points of the valley walls and farther from the receding bends; but beyond this point the analogy fails, for the trains of débris on the ice do not conform in detail to the courses of the currents of a winding stream, nor is there evidence of the rotatory motion that characterizes river water. Furthermore, the glacier is readily fractured, as the numerous gaping crevices on many glaciers show. The crevasses are sometimes longitudinal, sometimes transverse, and sometimes oblique. In the case of Arctic glaciers, longitudinal crevassing is especially conspicuous.

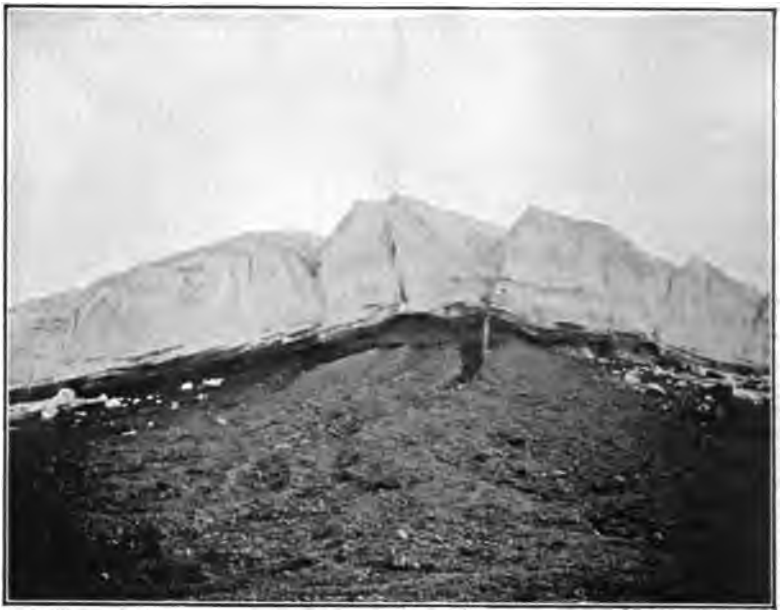


Fig. 238.—Cracking of glacier due to change in grade of bed. A North Greenland glacier overriding a mound of moraine-stuff.

Crevasses appear to be developed wherever there is appreciable tension, and the causes of this tension are many. An obvious cause is an abrupt increase of gradient in the bed (Fig. 238). If the change of gradient be considerable, an ice-fall or cascade results, and the ice may be greatly riven (Fig. 228). Below the cascade, the surface may bristle with wedges and pinnacles of ice (séracs, Fig. 239). Transverse crevices at the margin sometimes appear to be the result of the tension developed on a curve. Oblique crevices on the surface near the sides are commonly ascribed to the tension between the faster-moving center and the slower-moving margins, and in like manner crevasses that rise obliquely from the bottoms are attributed to the tension between the faster-moving portions above and the slower-moving portions below. All these crevasses indicate strains to which a liquid, whose pressures are equal in all directions, does not offer a close analogy. Longitudinal crevasses may affect both the river-like part of a glacier and its deploying end, and are the result of tension developed by movement within the ice itself, to which, again, rivers offer no analogy. Somewhat similar cracks develop in the outer crust of asphalt, when a mass of it is allowed to stand and spread; but in this case there is evaporation of the volatile ingredients, giving to the outer part relative rigidity and brittleness, while the inner part remains more fluent. The analogy is therefore not perfect and probably not really illustrative. The crevices may be narrow or wide, and both narrow and wide may be found in the same glacier. The narrow crevices that never open much are the most significant, as they show that very little stretching is needed to satisfy the tension. The opening of a gaping crevice is sometimes the work of weeks, and, in the slow-moving glaciers of high latitudes, sometimes the work of successive seasons. All this shows that the glacier is a very brittle body, incapable of resisting even very moderate strains brought to bear upon it very slowly. Had the ice even moderate duc-

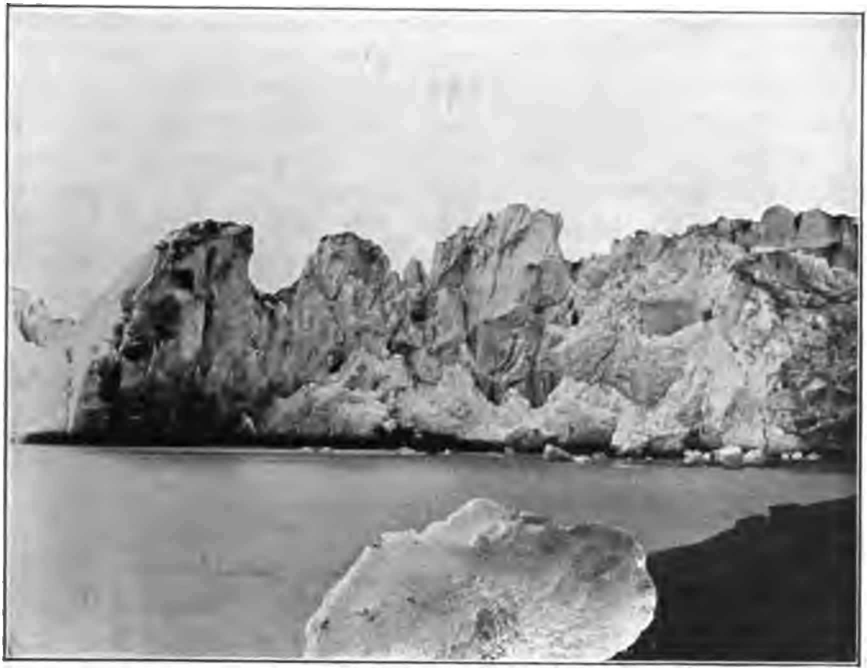


FIG. 239.—Séracs of glacier. (Reid.)

tility, it would adapt itself to tension brought to bear upon it so slowly as are many of the tensions which produce crevassing. In its behavior under tension therefore a glacier is notably unlike a river.

SURFACE FEATURES.

Topography. Many of the minor irregularities of the surface of a glacier are the result of crevassing. After the ice is crevassed, the sun's rays and the air which has been warmed by them penetrate the openings and melt the ice. The melting is most rapid at the top, and decreases downward. The result is that the sections of ice between adjacent crevasses are narrowed into wedges. If there be cross-crevassing, as is common, points instead of wedges result. As the sort of surface shown in Fig. 239 develops, any débris which was on the ice slides into the crevices, and the upper surface becomes clean.

Where ice is crevassed transversely, and where melting is not rapid, the crevasses may close as the ice moves forward, and the regelation of adjoining faces heals the rents in the surface. Even in this case, however, the surface is likely to be more or less undulating because of the waste on the sides of the crevices before they are closed. After regelation, surface ablation tends to obliterate the protuberances.

The topography of the surface of the ice is affected by other conditions. All parts of the surface of the ice are not equally compact, and the least compact portions melt most rapidly, giving rise to depressions, while the more solid parts occasion protuberances. Both depressions and protuberances may be regular or irregular in form (Figs. 240 and 241). Undulations of the bed often show themselves in the surface of the ice as suggested by Fig. 242. In such cases, ponds or lakelets sometimes accumulate on the surface of the ice. The topography of the ice in such cases seems to show that the ice is forced up slope.

Surface moraines.—The surface of a glacier is often affected by débris of one sort or another, and this also influences its topography. The débris is sometimes disposed in the form of belts or *moraines* (Figs. 237, 243). The surface moraines may be *lateral*, *medial*, or *terminal*. A *lateral moraine* is any considerable accumulation of débris in a belt on the side of a glacier. A *medial moraine* is a similar accumulation at some distance from the margins, but not necessarily in or near the

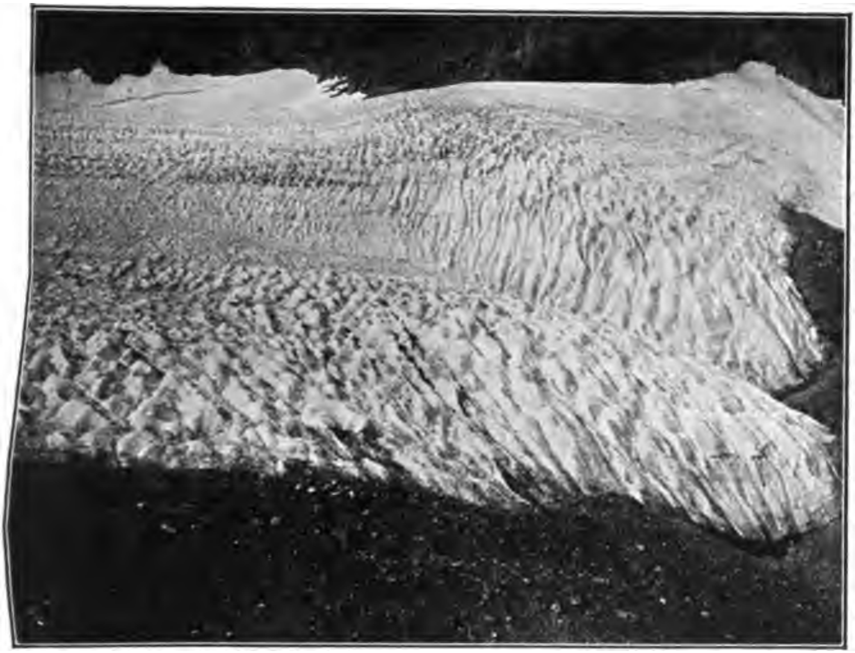


Fig. 240.—End of Mount Dana glacier, Cal. Shows irregularities of surface due to crevassing farther up the glacier, and to unequal melting.

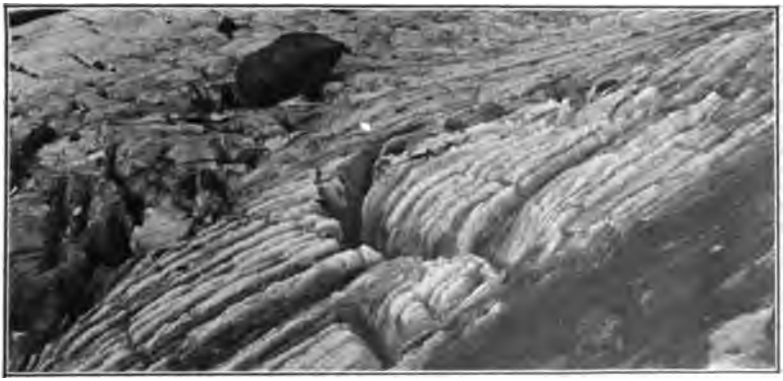


Fig. 241.—Shows irregularities due to unequal melting of veined ice. End of small glacier south of Forno hut, Engadine, Switzerland. (Reid.)

middle. There may be several medial moraines on one glacier, in which case some of them may be far from the center. In alpine glaciers, the *surface terminal* moraine is less well-defined; in polar glaciers it often connects two lateral moraines, making a loop roughly concentric with the terminus of the glacier.

Besides the surface moraines, which represent belted aggregations of débris, there may be scattered bowlders and bits of rock of various sizes on the ice, and, in addition to the coarse material, there is often some dust which has been blown upon the ice.

Relief due to surface débris.—The débris on the ice affects its topography by influencing the melting of the subjacent and adjacent ice. The rock débris absorbs heat more readily than the ice. A small and thin piece of stone lying on the ice is warmed through by the sun's rays, and, melting the ice beneath, sinks, just as a piece of black cloth on snow will sink because of the increased melting beneath it. Though a good absorber of heat, rock is a poor conductor, and so the lower surface of a large mass of stone is not notably warmed. The ice beneath it is protected from the direct rays of the sun, and is therefore melted more slowly than that around it. The result is that the bowlder presently stands on a protuberance of ice (Fig. 244). When its pedestal becomes high, the oblique rays of the sun and the warm air surrounding it cause it to waste away, and the capping bowlder falls. In high latitudes,

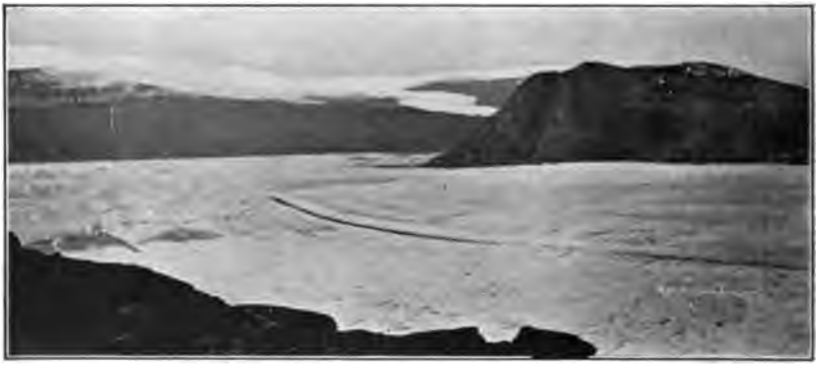


FIG. 242.—Irregular surface due to uneven bottom. Bowdoin glacier, Inglefield Gulf, North Greenland. The dark patches near the left margin of the glacier are lakelets in basins produced by the upward bending of the ice as it overrides an elevation in its bed. The figure also shows a depressed medial moraine.

the great obliquity of the rays sometimes allows them to strike under isolated bowlders. In this case, they are warmed from below, and thus aid rather than hinder the melting of the ice.

The same principles apply to the moraines. A thin bowlder moraine in high latitudes is sometimes sunk below the surface (Fig. 242). Usually, however, a medial moraine protects the ice beneath from melting, and occasions the development of a ridge of ice beneath itself. As the ice on either side is then lowered by ablation, the moraine matter of the medial belt tends to slide down on either hand. The same is true of the lateral moraines. So far does this spreading go, that in some cases the lower end of a glacier is completely covered with the débris which has spread from the medial and lateral moraines. Examples of this may be seen in almost any region of abundant, long, alpine glaciers.

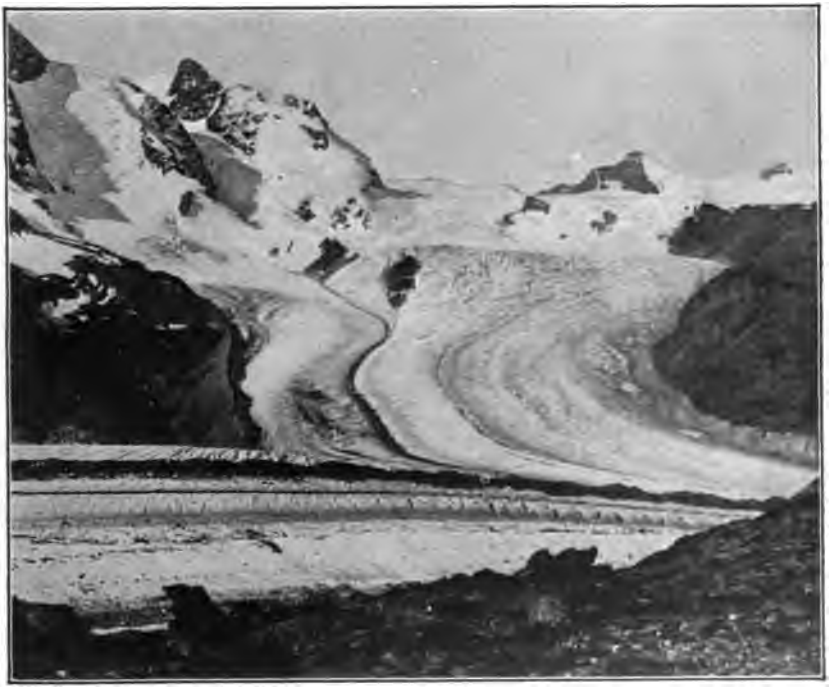


Fig. 243.—A Swiss glacier, showing surface moraines, characteristic profile, etc.

Dust-wells.—The wind-blown dust sometimes gives rise to peculiar topographic features of small size. The dust is not distributed by the wind with absolute equality, and the surface drainage of the ice tends to aggregate it. Every dust particle acts like a small stone, and where aggregations of dust occur, they melt their way down into the ice, developing holes or "dust-wells" (Fig. 245). These wells rarely reach a depth of more than a few inches, but they may be so numerous that the pedestrian is obliged to watch his steps. This is especially true near the edge of the large ice-caps. It is evident that the depth of these dust wells must be slight, for so soon as they are deep enough to cut off the sun's rays from the dust at the bottom, the deepening ceases. Other things being equal, they are deeper in low latitudes than in high.

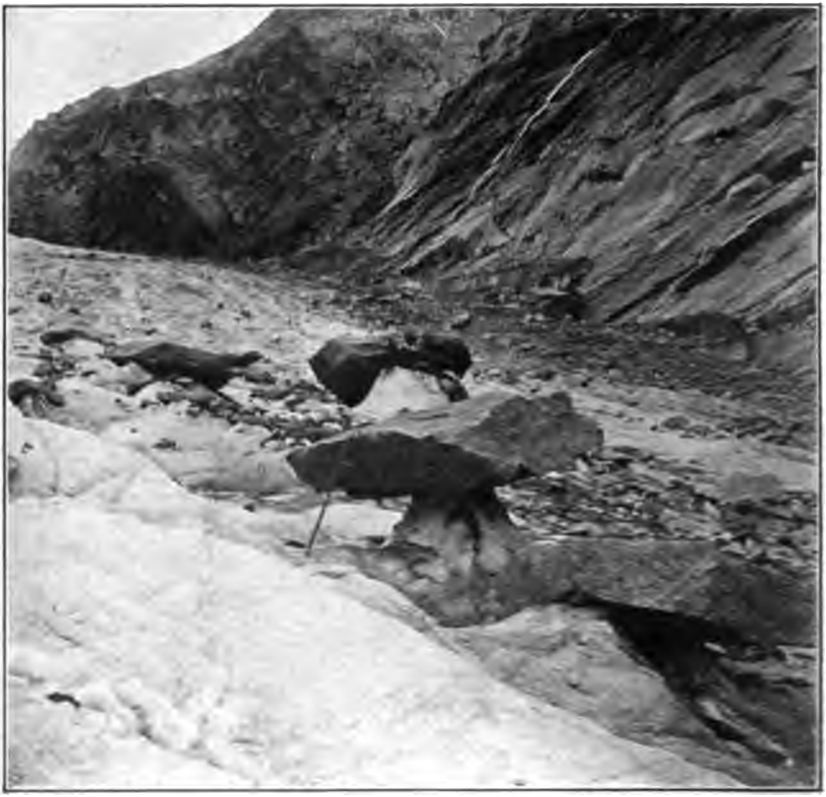


FIG. 244.—Bowlder on ice pinnacle. Forno glacier, Switzerland. (Reid.)

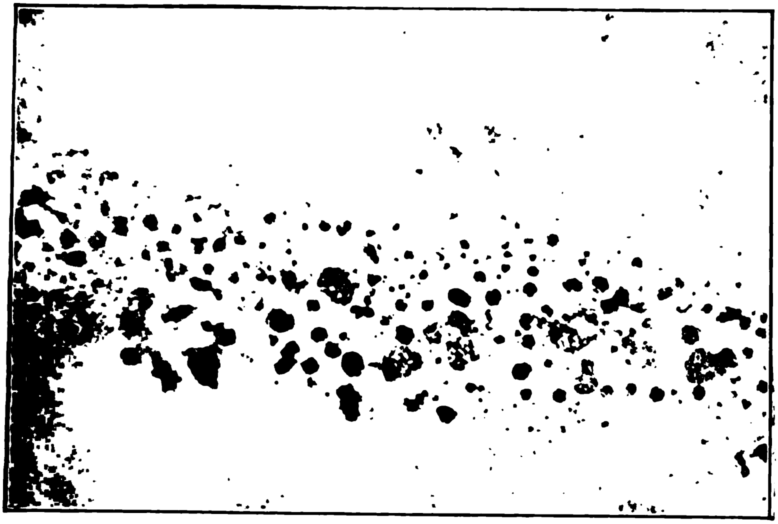
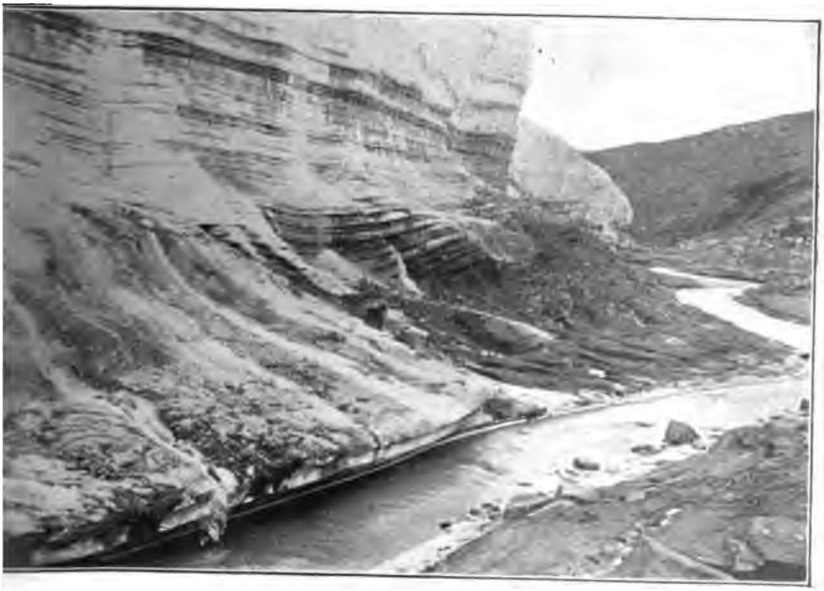


FIG. 245.—Dust-wells. Igloodahomyne glacier, North Greenland.



IG. 246.—Disposition of débris in ice. North Greenland glacier. (Libbey.)

Débris below the surface.—The lower part of a glacier, as well as the upper, carries rock débris. This débris is sometimes so abundant, especially near the ends and edges of the ice, that it is difficult to locate the bottom of the glacier; for between the moving ice which is full of débris, and the stationary débris which is full of ice, there seems to be a nearly complete gradation. The debris in the lower part of arctic glaciers, and to some extent of others, is often disposed in thin sheets sandwiched in between layers of clean ice. These débris sheets are often numerous and usually discontinuous, though groups of such sheets often persist for considerable distances. Débris also occurs to some extent in the ice well above its base. It is sometimes in belts, as seen in section, and sometimes in bunches. These various relations are illustrated by Figs. 227, 229, and 246–249.

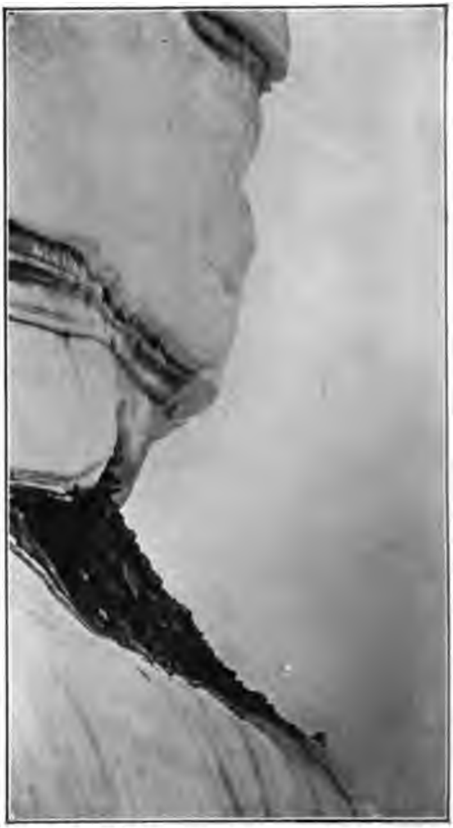


Fig. 247.—Profile of the lower part of the lateral margin of a glacier. Southeast side of McCormick Bay, North Greenland.

Another characteristic of the basal debris-laden part of some glaciers is the *foliation of the ice* (Figs. 248, 249, etc). This is especially well shown in the arctic glaciers, the ends and sides of which have steep or vertical faces. The foliation is best developed in the débris zone, though often shown above. The foliation is sometimes minute, consisting of layers of clean ice, an inch or less in thickness, separated by mere films of earthy matter. In extreme cases there are a score or more of laminæ within a foot. Locally, and especially where débris is abundant, the laminæ are much contorted. This is seen both in section (Figs. 248 and 249) and on the surface (Fig. 250).

TEMPERATURE, WASTE, AND DRAINAGE.

The temperature of glacier ice may range downward from the freezing point of water much as other solid portions of the earth's surface, but it has a fixed upper limit at 32° Fahr. (0° C.) because all the heat it receives tending to raise its temperature above that point, is converted into the latent form by the melting of the ice. The range of temperature is greatest at the surface, where it varies from 32° in the summer, to the coldest temperature of the region where the ice occurs. Beneath the surface the range of temperature is more restricted, and increasingly so with increasing depth.

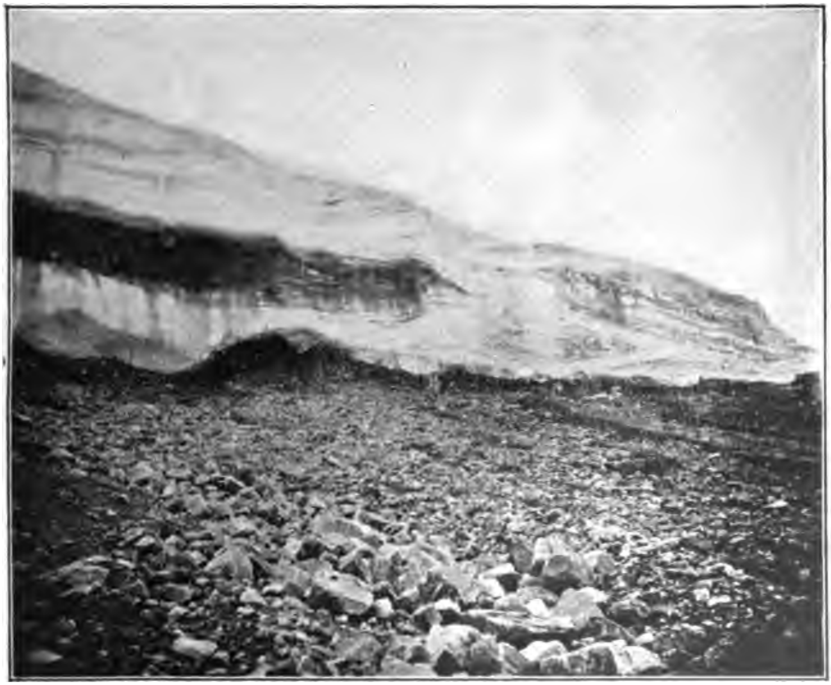


Fig. 248.—Side view of end of glacier. Southeast side of McCormick Bay, North Greenland. Shows structure of ice as well as position of débris.

The variation of temperature at the surface is due primarily to the varying temperature of the air. During the cold season, a wave of low temperature (the *winter* wave), starting at the surface, penetrates the ice, and during the warm season a wave of higher temperature (the *summer* wave) takes the same course. The day and night waves and other minor variables are, for present purposes, negligible.

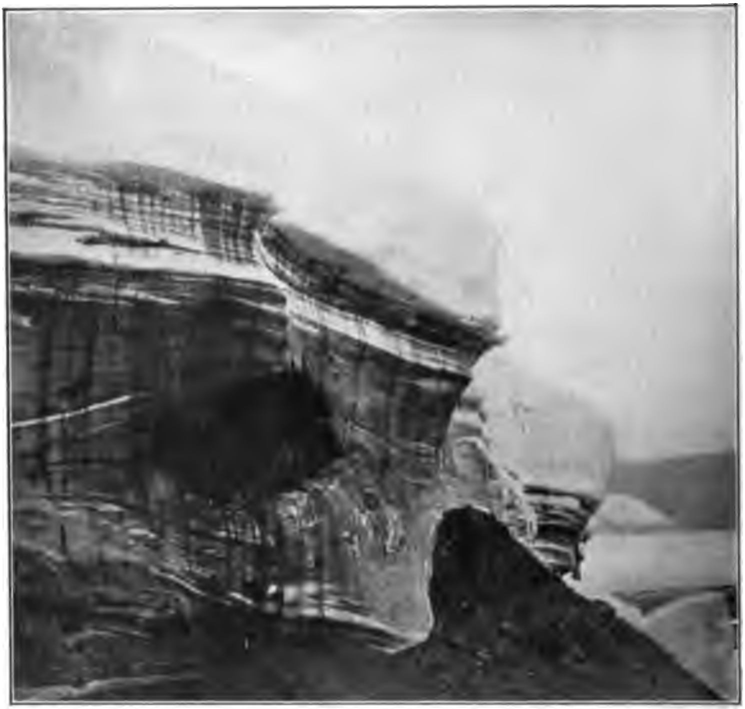
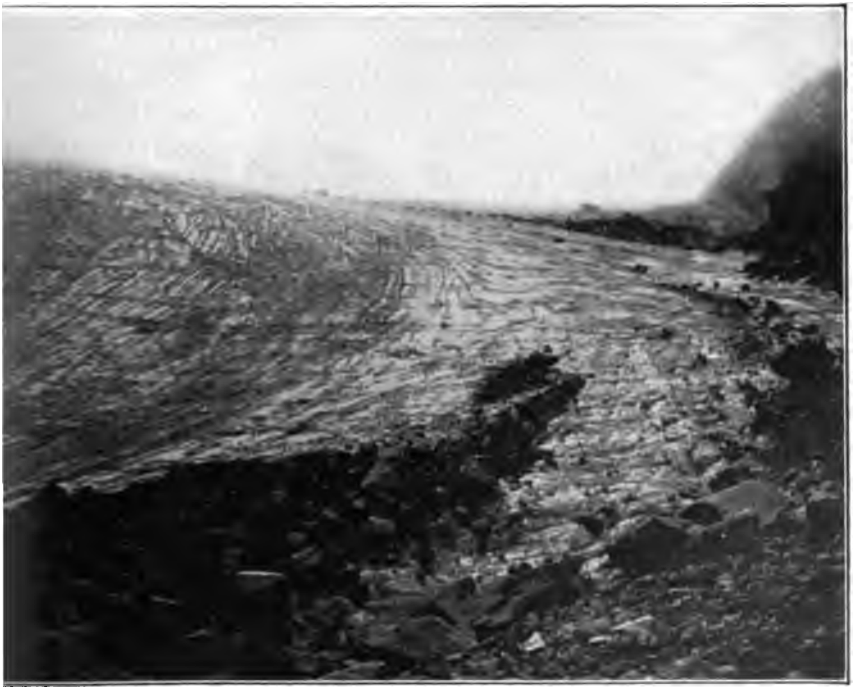


Fig. 249.—Side view of a North Greenland glacier (East glacier), showing position of debris and structure of the ice.

The winter wave.—There are but few observations on the internal temperatures of glaciers during the winter season, but it seems certain that the winter wave diminishes rapidly downward and dies out below, much as does the winter wave which affects land surfaces not covered with ice. Conduction alone considered, the temperature of the ice where the cold wave dies out, should correspond, approximately, to the mean annual temperature of the region, provided that temperature is below the melting point of ice.

Assuming that in the high altitudes and high latitudes where glaciers ound, the temperature of the surface may average about $-12°$ Fahr. out $-25°$ C.) for the winter half of the year, which is about the case north Greenland, Spitzbergen, and Franz Josef Land, and that the nductivity of the ice in the C. G. S.[1] system is .005, the temperature



G. 250.—Contorted lamination shown at the surface. A small glacier south of Forno hut, Engadine, Switzerland. (Reid.)

ould be lowered appreciably only about 40 feet below the surface at e close of the winter period, conduction only being considered. How r the internal temperature may be influenced by air forced through e ice by winds and by variations of the barometer is not known and nnot well be estimated. The wave of low temperature descending om the surface in winter would probably become inappreciable fore reaching a depth of 60 feet. At this depth the temperature ould be about 15° Fahr.—the mean annual temperature of the region.

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[1] Centimeter-gramme-second system. The rate of conductivity has not been ry accurately determined.

The summer wave.—The warm wave follows the analogy of the summer wave of ice-free land surfaces much less closely. This is because of the low melting temperature of ice as compared with other forms of solid earth-matter. On this account the summer wave is bifold. The one part travels downward by conduction, the other by the descent of water; the one has to do primarily with the temperature before the melting-point of ice is reached; the other, with the temperature after that point is reached; the first conforms measurably to the warm wave affecting other solid earth-matter, while the second is governed by special laws. After the surface portion of the ice is brought to the melting temperature, the additional heat which it receives melts the ice and is transformed from *sensible* into *potential* heat. Ice charged with water is *potentially*, but not *sensibly*, warmer than ice which has just reached the melting temperature.

The warm wave of conduction dies out below like the cold wave. The warm wave descending by the flow of water stops where the freezing temperature of water is reached. In regions where the average temperature is below freezing, the water-wave does not descend so far as the wave of conduction, since the latter descends below the zone where the melting temperature is found.

The foregoing considerations warrant the generalization that glaciers normally consist of two zones (1) an outer or upper zone of fluctuating temperature, and (2) an under zone of nearly constant temperature. The under zone obviously does not exist where the thickness of the ice is less than the thickness of the zone of fluctuating temperature. This may be the case in very thin glaciers in very cold regions, and in the thin ends and edges of all glaciers.

The temperature of the bottom.—The internal heat of the earth is slowly conducted to the base of a glacier where it melts the ice at the estimated average rate of about one-fourth of an inch per year. The temperature of melting is a little below 32° Fahr. since pressure lowers the melting-point at the rate of .0133° Fahr. (.0075° C.) for one atmosphere of pressure. At the bottom of a mile of ice therefore the melting temperature is about 30.2° Fahr. (–1° C.) It is probable that in all thick glaciers the temperature of the bottom is constantly maintained at the melting-point. This may be indicated by the streams which issue from beneath glaciers during the winter, though this criterion is hardly decisive since the issuing waters may be derived partly or wholly from the rock beneath. In glaciers or in parts of glaciers so thin as to lie wholly within the zone of fluctuating temperature, the temperature of the bottom is obviously not constant.

Temperature of the interior of the ice.—The variation of temperature of the surface of a glacier has already been shown to lie between a maximum of 32° Fahr. and the minimum temperature of the region where the glacier occurs. Lower, in the zone of fluctuating temperature, the variation is less, and where the zone of fluctuating temperature passes into the zone of constant temperature, variation ceases. The thickness of the zone of fluctuating temperature varies with the temperature of the region where the glacier occurs, being greatest where the winters are coldest. In the case of all glaciers except thin ones in very cold regions, the temperatures within the zone of constant temperature range from the mean annual temperature of the region at the top of the zone (provided this is not above the melting-point of ice at this depth) to the melting temperature of the ice at the bottom. Within these limits the range may be great or slight.

If we consider only the effects of the external seasonal temperatures and the internal heat of the earth, it appears that all the ice in the zone of constant temperature in the lower end of a typical alpine glacier should have a constant melting temperature, for the average temperature of regions where the ends of such glaciers occur is usually above 32° Fahr., and this determines a temperature of 32° Fahr. (or a little less) at the top of the zone, while a melting temperature is maintained at the bottom by the earth's interior heat. In thin glaciers of very cold regions, where the zone of constant temperature has relatively slight thickness, the low temperature descending from the surface may so far overcome the effect of internal heat as to keep the bottom of the ice at a freezing temperature. In all other cases, the ice at the bottom of the under zone has a melting temperature, while that above is probably colder.

In the higher altitudes and in the polar latitudes where glaciers are chiefly generated, the mean annual temperature of the surface is usually below the melting-point of ice. Here the temperature of the ice between the top and bottom of the zone of constant temperature must, on the average, be below the melting-point, unless heat enough is generated in the interior of the ice to offset the effect of the temperature above. For example, where the mean annual temperature is 20°

Fahr. or lower, as in middle Greenland, Spitzbergen, and Franz Josef Land, and at certain high altitudes in more southerly latitudes, the mean temperature in the zone of constant temperature should range from 20° Fahr. at the top to 32° Fahr. (or a little less) below; i.e., it should average about 6° below the melting-point. Under these conditions, all the ice in the zone of constant temperature, except that at its bottom, must be permanently below the melting-point, but it is perhaps worthy of especial note that much of it is but little below. In alpine glaciers the part of the ice affected by this constant low temperature (below freezing) is presumed to be chiefly that which lies beneath the snow-fields. In polar glaciers the low temperature probably prevails beneath the surface, not only throughout the great ice-caps, but also in the marginal glaciers which descend from them.

From these theoretical considerations we may deduce the generalization that in the zone of constant temperature *within the area of glacial growth*, the temperature of the ice is generally *below* the melting-point, while *within the area of wastage*, the temperature of the corresponding zone is generally *at* the melting-point.

Compression and friction as causes of heat.—The foregoing conclusions are somewhat modified by dynamic sources of heat. The compression arising from gravity, and the friction developed where there is motion, are causes of heat. Since friction occurs only when motion takes place, the heat which it generates is secondary and may, for present purposes, be neglected. Compression not only lowers the melting-point slightly, but it *produces heat at the point of compression.* Where the ice is granular, the compression, due to weight, takes place at the contacts of the grains. At intermediate points the pressure tends to cause them to bulge, and this has the effect of lowering the temperature of the bulging points. If therefore the compression be considerable, the granules may be warmed to the melting-point where they press each other, while at other points their temperature may be lower. In this case melting will take place at the points of compression, and the moisture so produced will be transferred to the adjacent parts of the granule and immediately refrozen. Melting at the points of compression would result in some yielding of the mass, and in some shifting of the pressure to new points where compression and melting would again take place. Thus the melting, the refreezing, and the attendant movement might go on until the limits of the power of gravity in this direction were reached. From considerations already adduced, it appears that the temperature in some parts of every considerable body of ice must be such as to permit these changes. The heat due to depression and friction may modify the theoretical deductions drawn above from atmospheric and internal influences.

Summary.—If the foregoing generalizations be correct, (1) the surface of a glacier is likely to be melted during the summer, (2) its immediate bottom is slowly melting all the time (unless the thickness of the ice be less than the thickness of the zone of annual variation or of permanent freezing temperature); (3) its subsurface portion *in the zone of waste* is generally melting, owing to descending water, compression, and friction; while (4) its subsurface portion *in the zone of growth* is probably below the melting-point except as locally brought to that temperature by compression, friction, and descending water, and at the bottom by conduction from the rock beneath.

Movement under low temperature.—Glacier motion will not be discussed at this point, but one of the bearings of the preceding conclusions on glacier motion may be pointed out. Since there must be motion in the area of growth to supply the loss in the area of waste, the *fundamental* cause of motion must be operative in bodies of ice the mean temperature of which is below the melting-point, *unless the dynamic sources of heat are considerable.* This fundamental cause does not exclude the *coöperation* of causes that work only (1) at the melting temperature, or (2) where the ice is bathed with water, or (3) in the plane of contact between wet ice above and dry ice below. These may be *auxiliary* causes which abet the fundamental one in producing the more rapid movement of warm seasons, or in bringing about the especially rapid movement in situations where there is abundant water, or in inducing the shearing which is such a remarkable feature of arctic glaciers.

Evaporation.—The ice wastes by evaporation as well as by melting, and while the former process is far less important than the latter, its results are probably larger than is commonly apprehended. One of the most remarkable features of some of the deposits of ancient glaciers is the slight evidence they afford of escaping waters. The most plausible explanation seems to lie in the supposition that the ice was largely wasted by evaporation. This conclusion finds support in many places in the presence of a mantle of fine silt over the drift, the silt being apparently composed of dust blown upon the ice. It is supposed to imply aridity in the region about the ice. If a sufficient mantle of dust were spread over the border zone of the ice, and if the air were very dry, nearly all the water melted on the surface of the ice might be held back by the dust-wells until the water was evaporated or absorbed.

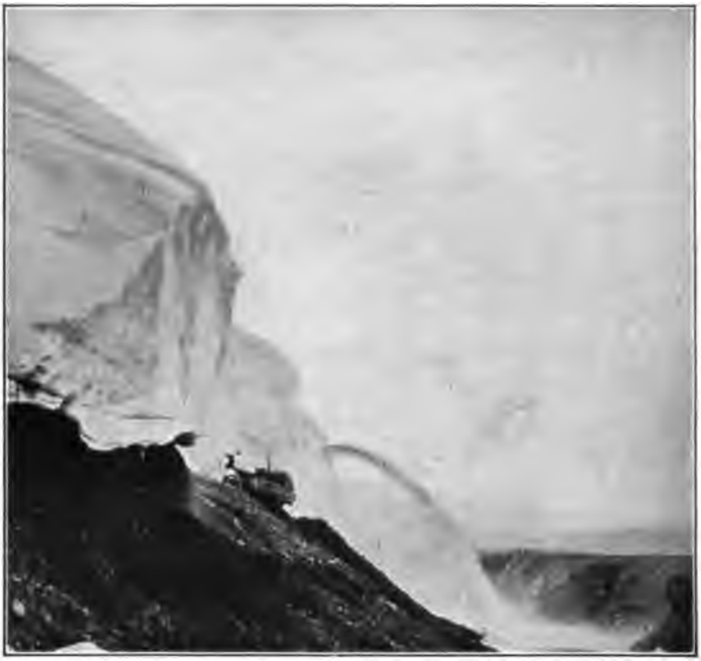

FIG. 251.—Spouting stream. Glacier south side of Olriks Bay, North Greenland.

Drainage.—Some of the water produced by surface melting forms little streams on the ice. Sooner or later they plunge into crevasses or over the sides and ends of the glacier. In the former case, they may melt or wear out well-like passages (moulins) in the ice, and even in the rock beneath. Much of the surface water sinks into the ice. Its ready penetration is aided by the "dust-wells" which mark the surface of many glaciers. In north Greenland wells which contain six or eight inches of water at the end of a warm day are often dry in the morning. The water has leaked out and passed to lower levels. From these and other harmonious observations it is inferred that the superficial part of a glacier at least is readily penetrated by water. The depth to which surface water penetrates is undetermined, but it doubtless varies greatly, not only in different glaciers, but in different parts of the same glacier, and in the same part at different times. Above the line of perennial snow there is little water either from melting or from rain, and hence relatively slight penetration. Below the line of perennial snow there is much melting and much rain, and here it is probable that the water sometimes, perhaps usually, penetrates to the bottom of the ice during the melting season, even independently of crevasses.

Once within the glacier, the course of the water is variable. Exceptionally it follows definite englacial channels, as shown by springs or streams issuing from the ice at some point above its bottom (Fig. 251). Oftener it descends or moves forward through the irregular openings which the accidents of motion have developed. If it reaches a level where the temperature is below its freezing-point, it congeals. Otherwise it remains in cavities or descends to the bottom. The water produced by melting within the glacier probably follows a similar course. So far as these waters descend to the bottom, they join those produced by basal melting and issue from the glacier with them. In alpine glaciers the waters beneath the ice often unite in a common stream in the axis of the valley, and hollow out a tunnel. Thus the Rhone is already a considerable stream where it issues from beneath the Rhone glacier. In the glaciers of high latitudes, sub-glacial tunnels are less common and the drainage is in streams along the sides of the glaciers or through the debris beneath and about them.

At the end of the glacier, all waters, whether they have been superglacial, englacial or subglacial, unite to bear away the silt, sand, gravel, and even small bowlders set free from the ice, and to spread them in belts along the border of the ice or in trains stretching down the valleys below. These are the most common of the glacio-fluvial deposits.

THE WORK OF GLACIERS.

Erosion and transportation.

The work accomplished by glaciers is distinctive, for while like rivers, they abrade the valleys through which they pass, carry forward the material which they remove from the surface, and wear, grind, and ultimately deposit it, and while their work therefore includes erosion, transportation, and deposition, their method is peculiar.

Getting load.—If the surface on which the snow-field which is to become a glacier accumulates be rough and covered with abundant rock débris, as such surfaces usually are, the glacier already has a basal load when its movement begins, for the snow covers, surrounds, and includes such loose blocks of rock as project above the general surface and envelops all projecting points of rock within its field. When the snow becomes ice and the ice begins to move, it carries forward the loose rock already imbedded in it, and tears off the weak points of the enveloped rock-projections. It may perhaps also move some of the soil and mantle rock of the original surface to which it is frozen. In addition to the *subglacial* load which the glacier thus has at the outset, there may be a surface load which has fallen on the snow or ice from cliffs above. This is especially true of mountain-valley glaciers. If this has been buried by snow and ice it is *englacial;* if it lies on the surface it is *superglacial.*

Once in movement, the ice carries away the debris to which it was originally attached, and at the same time gathers new load from the same area. The new load is acquired partly by the rasping effect of the rock-shod ice on its bed, and partly by its rending power which, under favorable circumstances, may quarry out considerable blocks of rock. This "plucking" process is at its best where the ice passes over cliffs of jointed rock or steep-sloped hills.

As the ice advances into new territory it acquires additional basal load, partly by rasping, partly by plucking, partly by freezing to it, and partly in other ways. One of these ways may be illustrated by the sequence of events when the end of a glacier advances on a very large bowlder. As the ice approaches it, the reflection of heat from it melts the adjacent edge of the ice, making a slight reentrant. With farther advance, the ice closes in against and around the bowlder, and finally carries it along in the bottom of the moving mass. In some cases, especially when its advance is rapid, the ice may push débris in front of itself. Even where this is the case, the amount of material pushed forward is generally slight, partly because the extreme edge of the ice often fails to rest on the land in summer, when the movement is greatest, being melted from below by the heat of the surface over which it is spreading (see Fig. 252), and partly because the earth in front of the glacier is frozen during a large part of the year. In this condition, the earthy matter has greater resistance than the ice, and the latter rides over it. Superglacial material may be acquired during movement by the fall of débris from cliffs, or by the descent of avalanches.

Conditions influencing rate of erosion.—An obstructive attitude of the surface toward the movement of the ice is as necessary for effective erosion as the movement of the ice itself. Advancing over a flat surface, ice ordinarily inflicts but little wear, since there is little for it to get hold of. So slight is the abrasive power of ice under these conditions that it frequently overrides and buries the soil with more or less of its herbaceous vegetation. But while a certain measure of roughness of surface is favorable for glacial erosion, the topography may be so uneven as to seriously impede the ice. Erosion is probably at its maximum, so far as influenced by topography, when the roughness of the surface is such as to offer notable catchment for the basal ice, but not such as to impede its motion very seriously. The amount of relief favorable for the greatest erosion increases with increasing thickness of the ice.

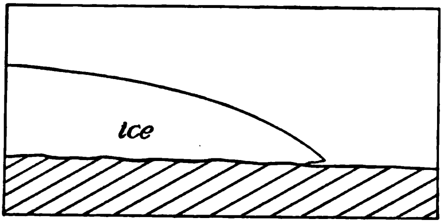

FIG. 252.—Diagram showing lack of contact of the edge of the icè with its bed.

Other conditions which influence erosion by ice are (1) the amount of loose or slightly attached débris on the surface, (2) the resistance of the rock, (3) the slope of the surface, (4) the thickness of the ice, (5) the rate of movement, and (6) the abundance and character of the débris which the ice has to work with. The effect of the first five of these conditions is evident. The effect of the last is less simple. Clean ice passing over a smooth surface of solid rock has little effect upon it; but a rock-shod glacier will abrade the same surface notably. The effect of this abrasion is shown in the grooves and scratches (*striæ*) which the stones in the bottom of the ice inflict on the surface of the rock over which they pass (Figs. 253, 255, and 256). At the same time the stones in the ice are themselves worn both by abrasion with the bottom, and with one another (Fig. 254). It does not follow, however, that the more material in the bottom of the ice the greater the erosion it effects; for with increase of débris there may be decrease of motion[1] and, beyond a certain point, the decrease of motion seriously interferes with the efficiency of erosion. When any considerable thickness of ice at the bottom of the glacier is full of débris, this loaded basal portion may

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[1] Russell. Jour. of Geol., Vol. III, p. 823.

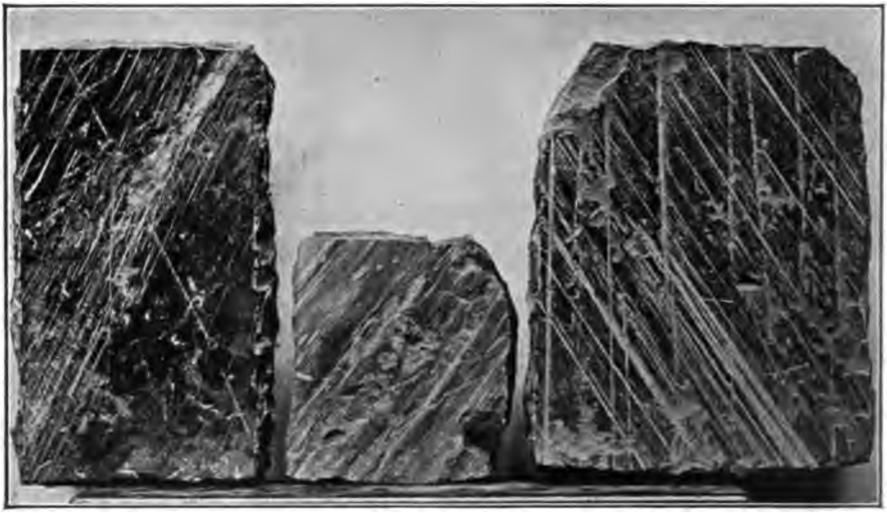


Fig. 253.—Glacial striæ and bruises. The block to the right shows two sets of striæ: that to the left shows the peculiar curved fractures known as *Chatter Marks*.

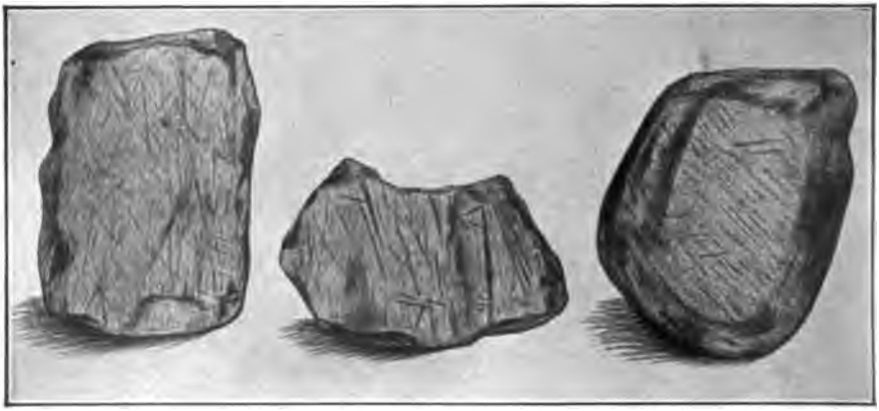


Fig. 254.—Bowlders showing glacial striation. (Drawn by Miss Matz.)

approach stagnancy, and the lower limit of considerable movement may lie between the loaded ice below and the relatively clean ice above. A moderate, but not an excessive load of débris is, therefore, favorable for great erosion. Something depends, too, on the character of the load. Coarse, hard, and angular débris is a more effective instrument of erosion than fine, soft, or rounded material. The adverse influence of the overloading of the ice on its motion has been likened to the stiffening of a viscous liquid by the addition of foreign matter, but it may better perhaps be referred to the destruction of the granular-crystalline continuity on which glacier motion probably depends.

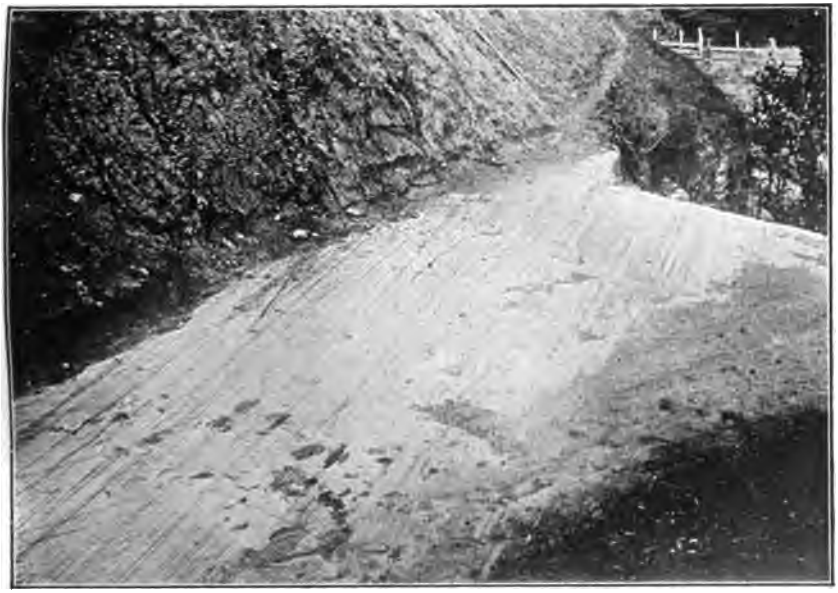


Fig. 255.—Striæ on bed rock, Kingston, Des Moines Co., Ia. (Iowa Geol. Surv.)

From the preceding statement, it is evident that erosion is not equally effective at all points beneath a glacier. So far as concerns the ice itself, erosion is not most effective at the end of a valley glacier, or at the edge of an ice sheet, for here the strength of movement is too slight and the load too great; nor is the most effective erosion at the source or near it, for though the ice may here be thick, the movement is slow and the load likely to be slight. Ice conditions only being considered, erosion is most effective somewhere between the source and the terminus, and probably much nearer the latter than the former. The conditions of the surface over which the ice passes may be such as to vary the place of greatest erosion widely. Thus in an Alpine glacier, erosion may be most effective at the *Bergschrund* because the slope here favors "plucking." Here, notable amphitheatres (*cirques*) are sometimes excavated. After the glacier disappears, the bottom of the cirque is often seen to contain rock basins (Fig. 257). Glacial cirques abound in mountains where glaciers once existed, but from which they have now disappeared.

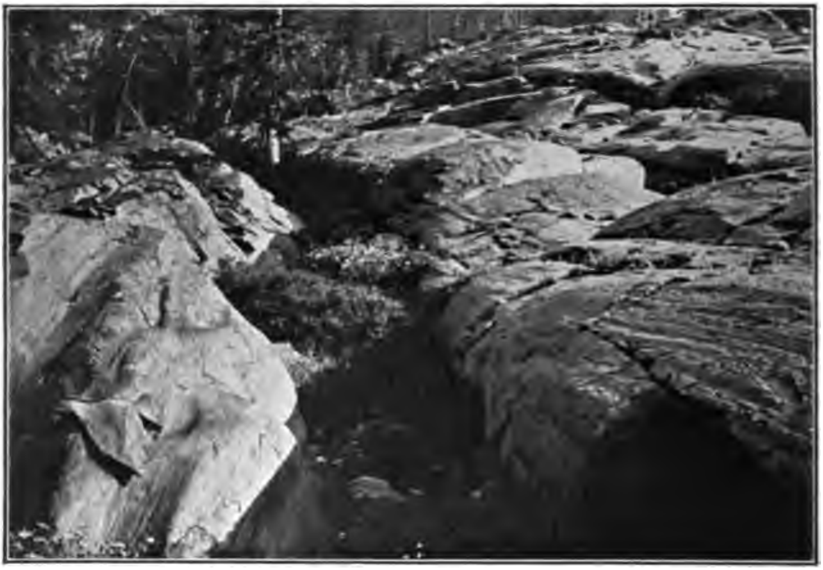


FIG. 256.—Striæ, grooves, etc., in a canyon tributary to Big Cottonwood Canyon, Wasatch Mountains. (Church.)

The cirques of the Bighorn mountains of Wyoming (Pl. XIX) are examples.

Summary.—In summary it may be said that rapidly moving ice of sufficient thickness to be working under goodly pressure, shod with a sufficient but not excessive quantity of hard-rock material, passing over incoherent or soft formations possessing a topography of sufficient relief to offer some resistance, and yet too little to retard seriously the progress of the ice, will erode most effectively.

Varied nature of glacial débris.—From its mode of erosion it will readily be seen that the bottom of a glacier may be charged with various sorts of material. There may be (1) bowlders which the ice has picked up from the surface, or which it has broken off from projecting points of rock over which it has passed; (2) smaller pieces of rock of the size of cobbles, pebbles, etc., either picked up by the ice from its bed or broken off from larger masses; (3) the fine products (rock-flour) produced by the grinding of the debris in the ice on the rock-bed over which it passes, and similar products resulting from the rubbing of stones in the ice against one another; and (4) sand, clay, soil, vegetation, etc., derived from the surface overridden. Thus the materials which the ice carries (*drift*) are of all grades of coarseness and fineness, from large bowlders to fine clay. The coarser material may be angular or round at the outset, and its form may be changed and its surface striated as it is moved forward. Whether one sort of material or another predominates, depends primarily on the nature of the surface overridden.

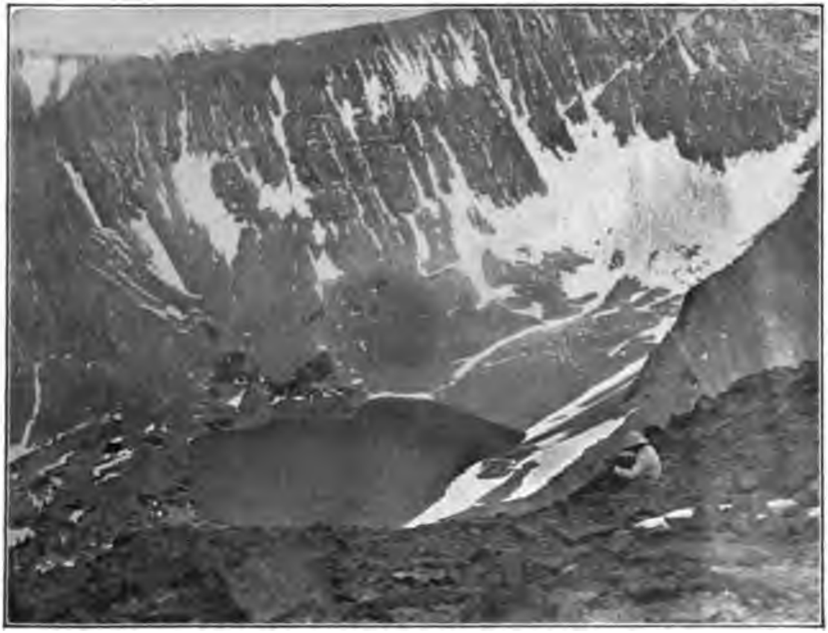


FIG. 257.—A glacial cirque. The lake occupies a rock basin, produced by glacier erosion. Head of Little Timber Creek, Montana.

The topographic effects of glacial erosion.—In passing through its valley, an alpine glacier deepens and widens its bottom and smooths its slopes up to the upper limit of the ice. It tends to change a V-shaped valley (Fig. 258) into a U-shaped one (Fig. 259). The change in topography at the upper limit of glaciation is often marked (Figs. 260 and 261).

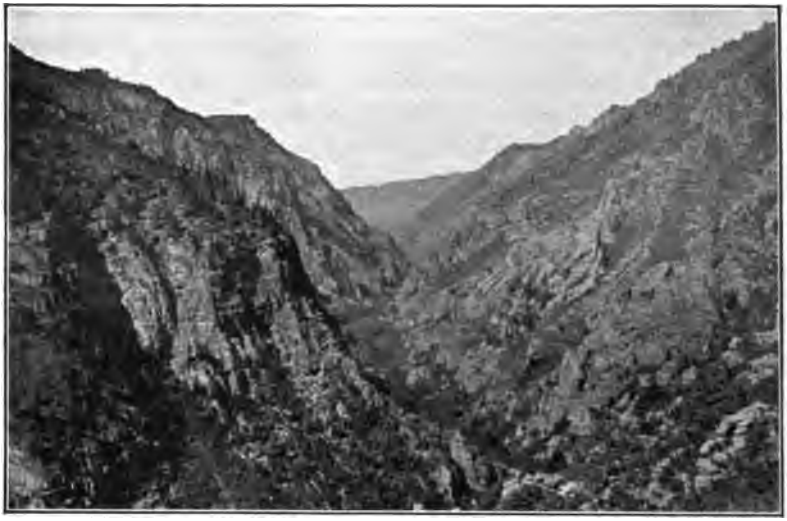


Fig. 258.—The Valley of the American Fork. A V-shaped non-glaciated valley in the Wasatch Mountains of Utah. Compare Fig. 259. (Church.)

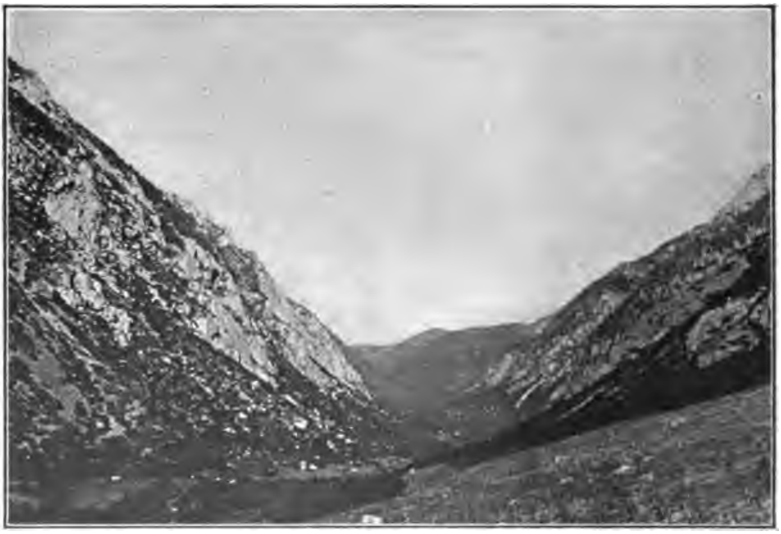


Fig. 259.—U-shaped valley resulting from glaciation. Little Cottonwood Canyon, Wasatch Mountains. (Church.)

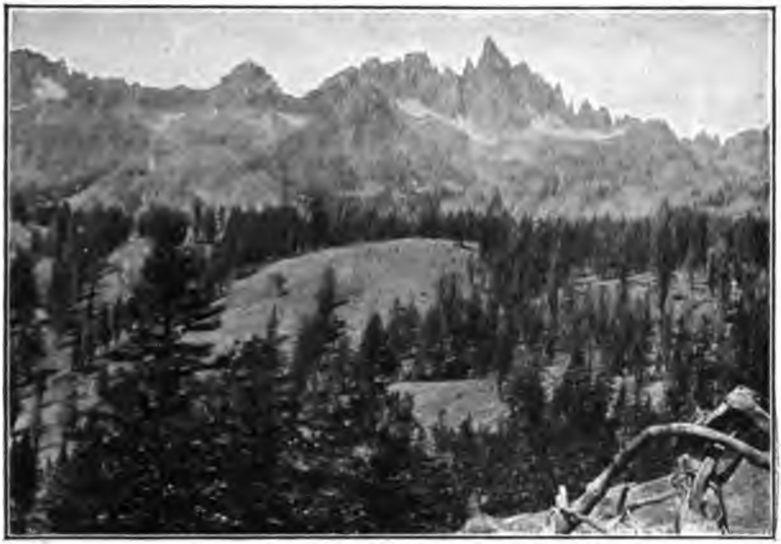


Fig. 260.—Contrast between glaciated topography below and non-glaciated topography above. The minarets in the Sierras, Cal.

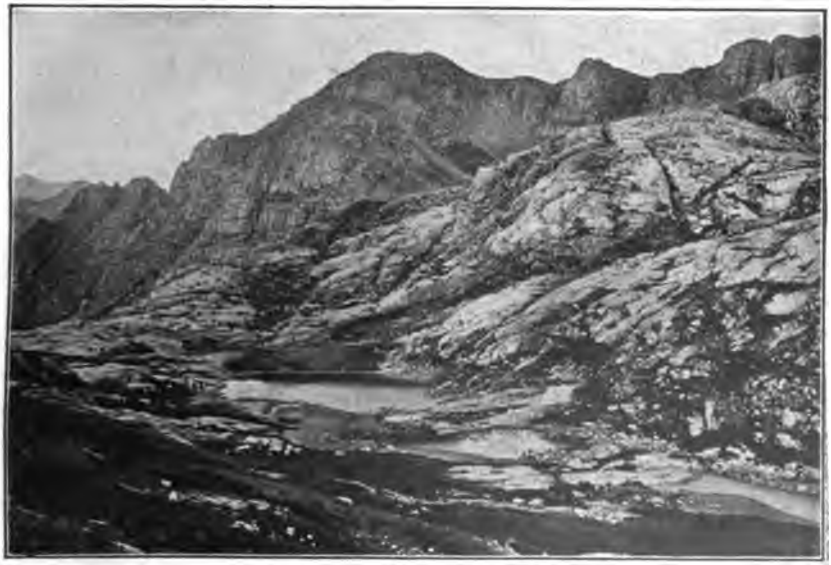


Fig. 261.—Contrast between glaciated topography below, and non-glaciated topography above. Needles Mountains, from slope west of Hidden Lake. (Cross, U. S. Geol. Surv.)

The deepening of a valley by glacial erosion may throw its tributaries out of topographic adjustment. Thus if a main valley is lowered 100 feet by glacial erosion while its tributary is not deepened, the lower end of the latter will be 100 feet above the former when the ice disappears. Such a valley is called a *hanging valley* (Figs. 262 and 263). Such valleys are of common occurrence in regions recently glaciated, but now ice-free. Examples are common in the western mountains of North America and elsewhere.

Ice-caps which overspread the surface irrespective of valleys and hills, tend to reduce the angularities of the surface. Hills and ridges are cut down and smoothed (Figs. 264 and 265); but since valleys parallel to the direction of movement are deepened at the same time, it is doubtful if the relief of the surface is commonly reduced by the erosion of an ice-cap.

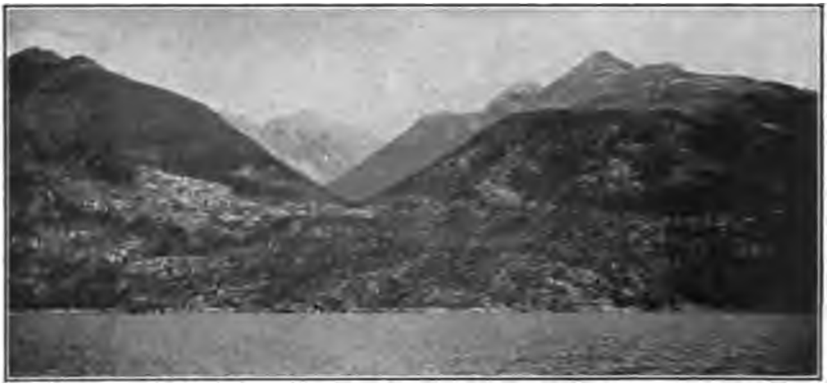


FIG. 262.—A hanging valley. East side of Lake Kootenai, B. C. All except the highest summits glaciated. (Atwood.)

Fiords.—A glacier descending into the head of a narrow bay may gouge out the bay to a very considerable depth, causing its head to recede. When the ice finally melts, the bay may be a fiord. Thus have arisen the glacial features of many of the fiords of high-latitude coasts, and many of the glaciers of those coasts are now making fiords (Fig. 266). Fiords also arise in other ways. Coasts indented by fiords are likely to be bordered by islands.

The positions in which débris is carried.—As a result of the methods by which a glacier becomes loaded, there are three positions in which the débris is carried: (1) the basal or subglacial, (2) the englacial, and (3) the superglacial. The material picked up or rubbed off from the surface over which the ice moves is normally carried forward in the base of the ice; while that which falls on the surface is usually carried in the form of surface moraines. In the former position the drift is basal; in the latter, superglacial. It is doubtful if much débris is moved along beneath (that is, strictly below the bottom of) the ice, though the movement of the latter would have a tendency to drag or urge along with it the loose material of its bed. If drift were carried forward in such positions, it would be strictly *subglacial.*

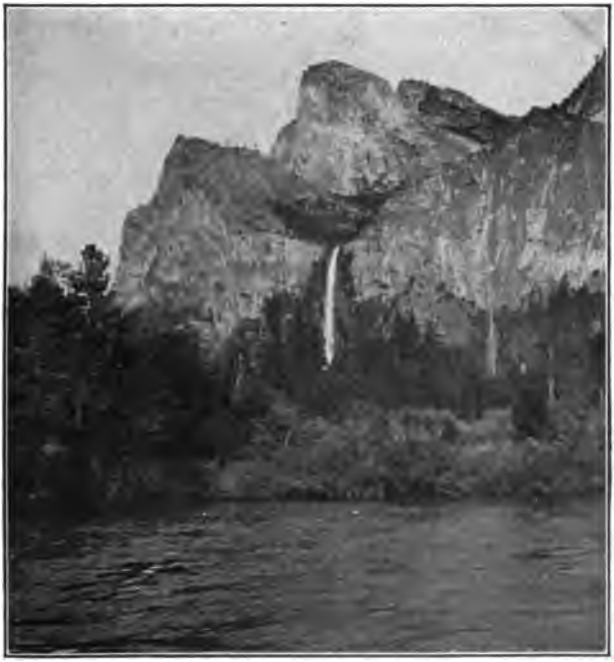


FIG. 263.—A hanging valley. The water falls (Bridal Veil) from a hanging valley. (Wineman.)

The basal load of a glacier is constantly being mixed with new accessions derived from ground over which the ice is passing, and this admixture tells the story of the work done by the bottom of the ice. The englacial and superglacial material, on the other hand, is normally borne from the place of origin to the place of deposition without such intermixture. It is a case of "local" versus "through" transportation.

Transfers of load.—While the origin of the load usually determines its position, exceptions and complications arise from the transfer of load from one position to another, and from the gradation of one horizon into another.

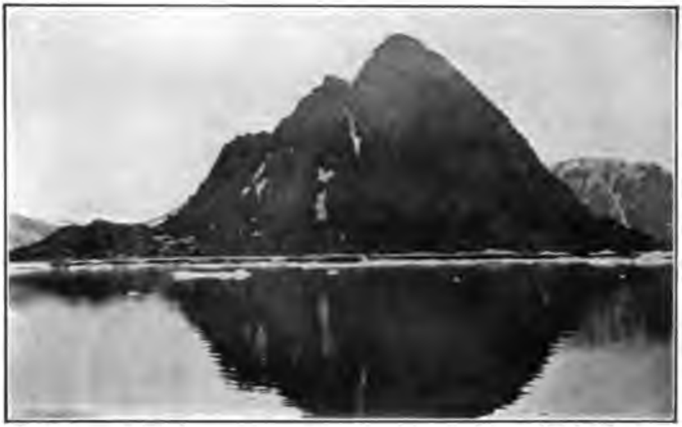


Fig. 264.—A non-glaciated hill. Dalrymple Island, North Greenland.

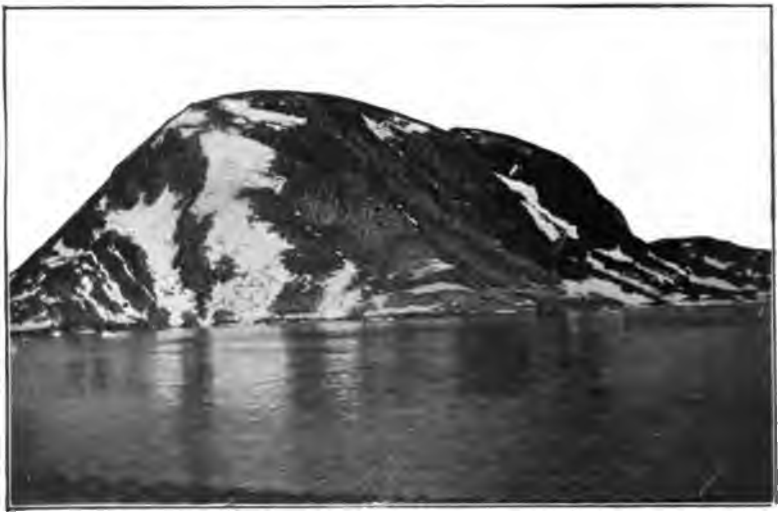


Fig. 265.—A glaciated hill. Southeastern Carey Island. About 30 miles west-northwest of Dalrymple Island (Fig. 265.)

Most of the débris gathered by ice is acquired at its bottom. While such material is basal *at the outset,* some of it may find itself above the bottom a little later. Thus when ice passes over a hill (Fig. 267) the bottom of the ice rends debris from the top of the hill. When it descends from one level to another there is a similar result (Fig. 268). To the lee of the hill the ice from either side may close in under that which came over the top, in which case the débris derived from the top of the hill by the bottom of the overriding ice will be well up in the ice. It has passed from an initial basal to a subsequent englacial position. The change does not usually involve an actual rise of the material,

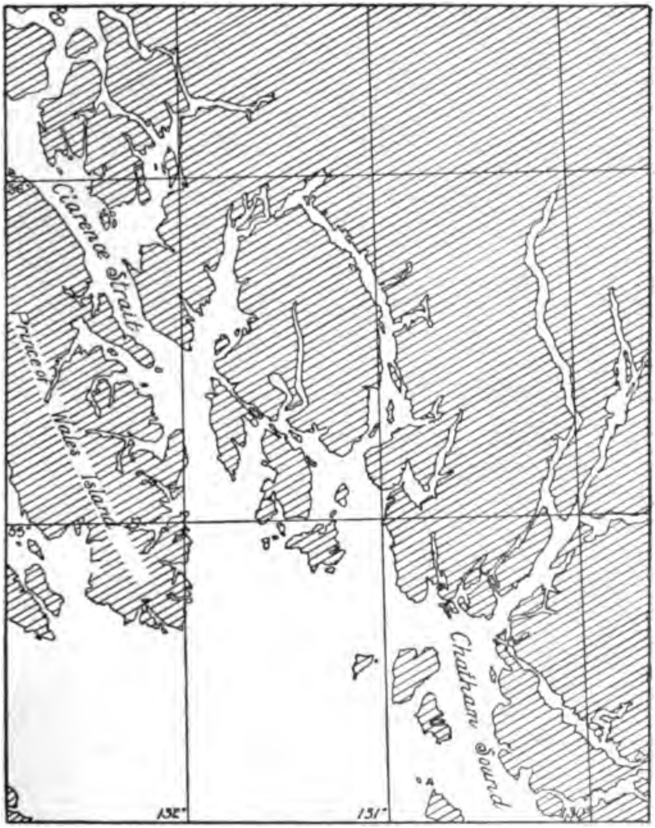


Fig. 266.—Alaskan fiords. The shaded areas represent land. (From charts of the C. & G. Surv.)

but rather a decline. If carried upward at all, the upward movement is temporary only, and incident to the passage of the ice over the hill, or to other local causes. The englacial débris may be little or much above the basal zone according to the height of the elevation overridden.

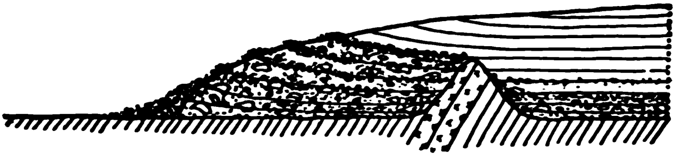


FIG. 267.—Diagram to illustrate the taking of débris from a hill-top. It also illustrates how englacial débris may become superglacial as the result of surface ablation.

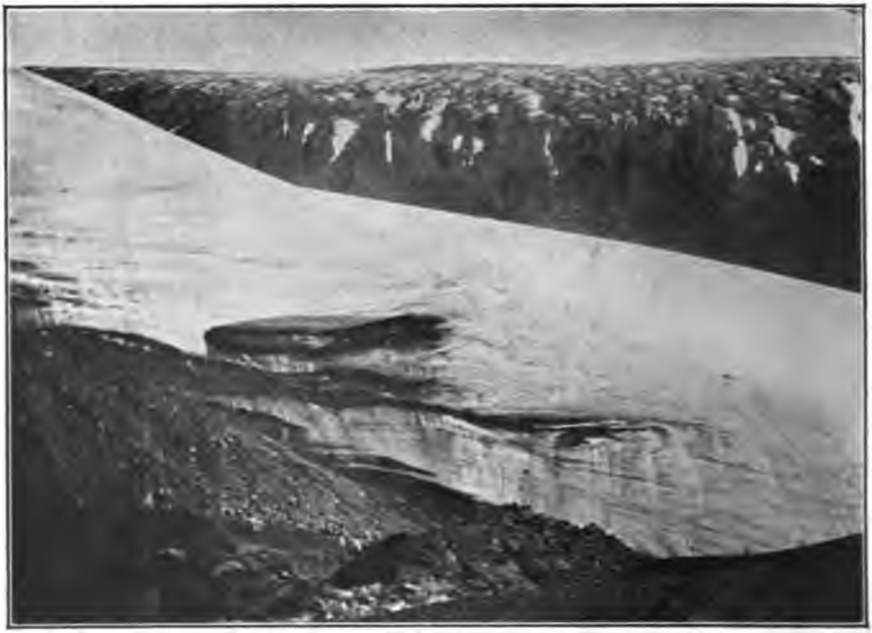


FIG. 268.—Taking débris from a protuberance of the bed.

Superglacial débris may obviously become englacial by falling into crevasses or by being carried down by descending waters. Either superglacial or englacial débris may become basal by the same means.

From their form and position, there is less ice-free land in immediate association with ice-caps than with valley glaciers. Furthermore, the ice-free land about the borders of an ice-cap is less likely to be in the form of cliffs above it. As a result, the surfaces of ice-caps are comparatively clean, except at their edges where the ice is thin.

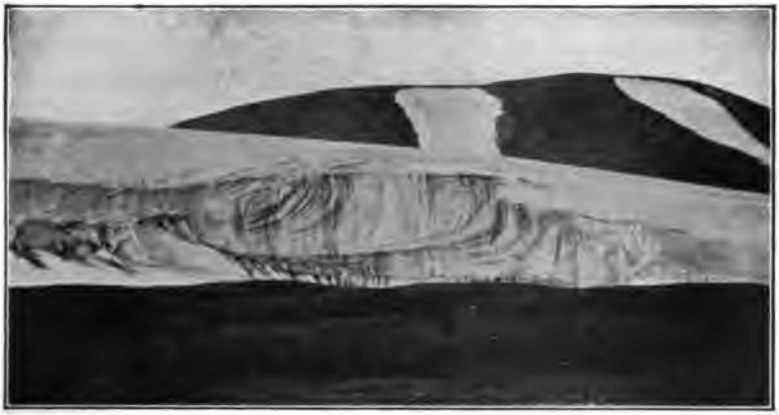


Fig. 269.—Side view of end of a glacier on the south side of Olriks Bay, North Greenland.

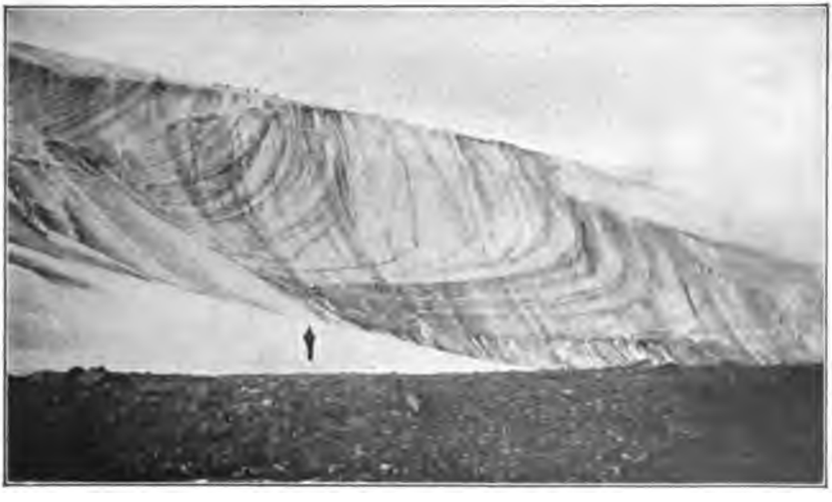


Fig. 270.—Closer view of a part of the ice shown in Fig. 269.

Englacial material may become superglacial by surface ablation. In this case the drift does not rise, but melting brings the surface of the ice down to its level. This occurs chiefly at the end or edge of the ice, where the surface melting is greatest. Englacial débris, especially that near the bottom, may also become basal by the melting of the bottom of the ice.

Englacial material plucked or rasped from an elevation over which the ice has passed is liable to be disposed in a longitudinal belt in the ice in the lee of the elevation itself. By surface ablation this material may reach the surface at some point below its source, and be disposed as a medial moraine. Such a moraine has an origin very different from that of a medial moraine formed by the junction of two lateral moraines of superglacial origin.

Much less in the natural order of things is the transfer of material from a basal to an englacial and from an englacial to a superglacial position by upward movement of the débris itself. Such transfer is remarkable because the specific gravity of rock is from two and a half to three times as great as that of ice, so that its normal tendency is to sink.

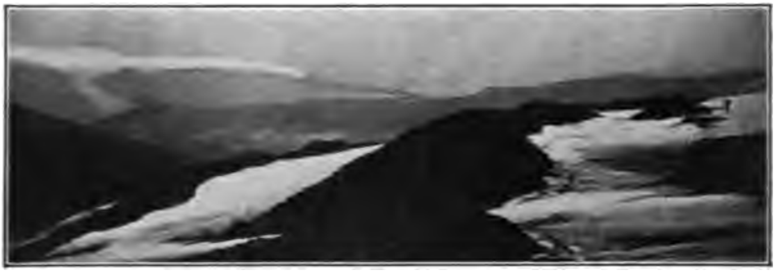


Fig. 271.—Surface terminal moraines due to upturning. Edge of the ice-sheet, North Greenland.

In arctic glaciers, and probably in others, some material which has been basal becomes englacial by being sheared forward over ice in front of it. So far as observed this takes place chiefly where the ice in front of the plane of shearing lies at a lower level than that behind, as where the surface of an upland falls off into a valley, or where a boss of rock shelters the ice in its lee from the thrust of the overriding ice (Fig. 268).

At the borders of arctic glaciers the lower layers are not infrequently upturned, as shown in Figs. 269 to 272. Where the layers turn up at the end of a glacier (Figs. 269 and 270), basal and englacial débris is carried to the surface by actual upward movement, and a terminal moraine or a series of terminal moraines sometimes aggregated *where the upturned layers of ice outcrop at the surface* (Fig. 271). That the material of these moraines was originally basal is abundantly demonstrated by the bruised and scratched condition of the bowlders and pebbles, and sometimes by the nature of the material itself. For example, in two cases in North Greenland where glaciers descend into the heads of shallow bays and move forward on their bottoms, moraines formed by the upturning of the layers were seen to contain *abundant molluscan shells derived from the bottom of the bay.* The upturning sometimes affects the side-edges of ice-tongues (Fig. 272) as well as their ends, and the material thus brought to the surface gives origin to lateral moraines altogether different in origin from the lateral moraines formed by the falling of debris upon the glaciers. Sometimes also there is an upturning of the ice along a longitudinal zone well back from the lateral margins (Fig. 273), and the material so borne to the surface in such a zone gives rise to a moraine resembling the medial moraine formed by the union of lateral moraines, but of wholly different origin.

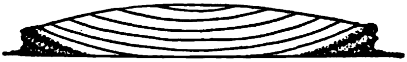


Fig. 272.—Diagram illustrating the upturning of the layers of ice at the end of an arctic glacier as seen in end-section. The bottom line represents sea level.

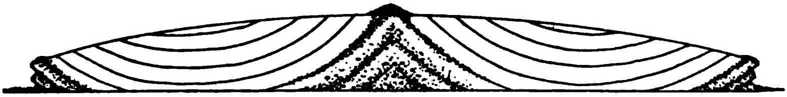


Fig. 273.—Diagram illustrating the same point as 272, where the structure is more complex. The bottom line of the figure represents sea level.

The phenomenon of upturning here referred to has been observed only at or near the terminus of the ice, and is perhaps due in most part to the resistance of frozen morainic or other material beneath and in front of the edge. To this should probably be added the effect of the increased rigidity of the ice at its borders, due to the low external temperature during the larger part of the year, while the interior, with its higher temperature, remains more fluent. But even this probably leaves the explanation inadequate. In not a few instances the upturning is associated with a notable *thickening* of the layers toward their edges (Fig. 274). This suggests that perhaps there is an exceptional growth of the granular crystals of the ice near the edge of the layers, owing to the penetration of the surface-waters which are much more abundant at the borders than elsewhere, and which in the arctic glaciers probably do not penetrate deeply before they reach a freezing temperature.

Wear of drift in transit.—Drift carried at the bottom of the ice is subject to notable wear. The materials in transportation abrade one another and are abraded by the bed over which they pass. Englacial drift is subject to less wear because it is commonly more scattered. Superglacial drift is worn little or none while it lies on the surface of the ice; but in so far as superglacial or englacial drift is derived from the basal load, it may show the same evidences of wear as the basal drift itself. Superglacial drift often reveals its history in this way.

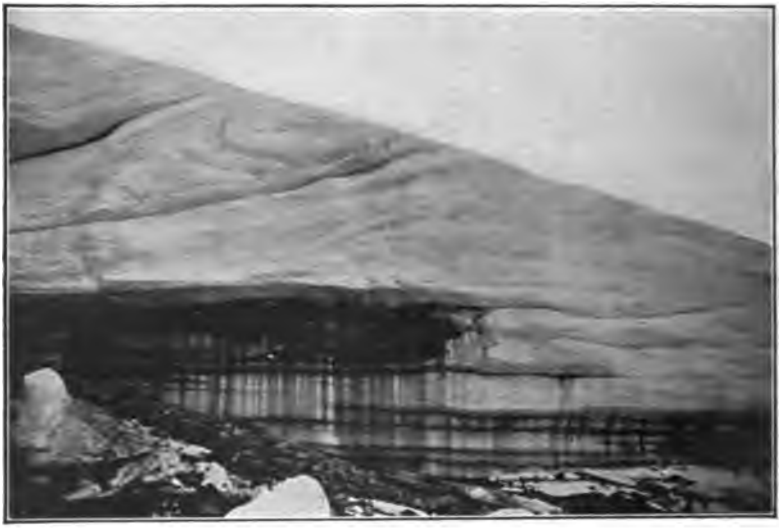


Fig. 274.—Thickening of the upturned layers of ice.

Deposition of the Drift.

1. **Beneath the body of the ice.**—During the advance of a glacier, deposition may take place both beneath the body of the ice and beneath its end and edges. Deposition beneath the body of the ice is liable to e place wherever the topography favors lodgment, or wherever the s overloaded. The topography favoring deposition is much the

same as that favoring erosion, but the two processes are not favored at the same point. Erosion is greatest on the "stoss" side of an obstruction (the side against which the ice advances), and deposition on the lee side. The ice is likely to be overloaded (1) just beyond a place where conditions have favored the gathering of a heavy load, and (2) where the ice is rapidly thinning. On the whole, however, the deposition of material beneath the main body of a glacier is much more than balanced by erosion in the same position.

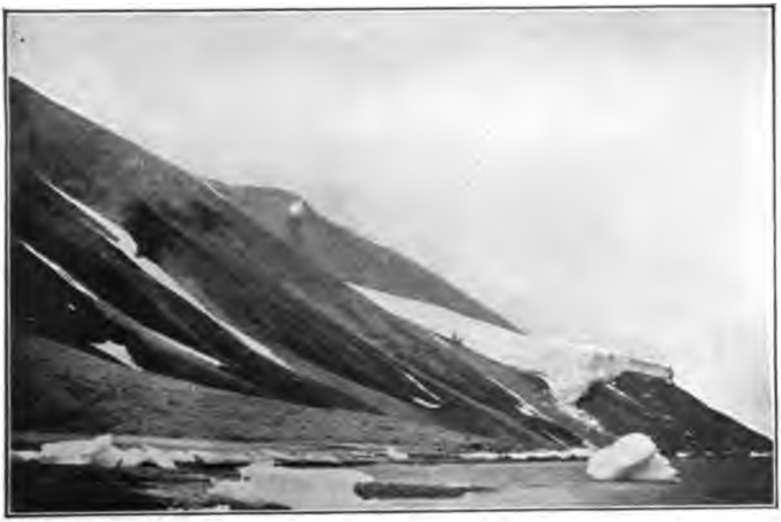


Fig. 275.—Glacier building an embankment. Southeast side of McCormick Bay, North Greenland.

2. **At ends and edges of glaciers.**—At and near the end of a glacier the conditions of deposition are somewhat different. Here deposition beneath the ice goes on faster than elsewhere, chiefly because of the more rapid melting and the more rapid thinning and weakening of the ice. If the end of the glacier be stationary in position, drift is being continually brought to it and left there, for though the *end* is stationary, the *ice* continues to move. If the glacier moves forward 500 feet per year, and if its end is melted at the same rate, all the debris in the 500 feet of ice which has been melted has been deposited, and all except that which has been washed away has been deposited at and

beneath the end of the glacier. If the end of the glacier is retreating, the retreat means that the waste at the end exceeds the forward movement. If the ice advances 300 feet per year, and is melted back 500 feet in the same time, all the débris carried by the 500 feet which has been melted has been deposited, and largely in the narrow zone (200 feet) from which the ice has receded. Even in this case, therefore, there is a notable tendency to marginal accumulation. If the end of the glacier is advancing 500 feet per year while it is being melted but 300 feet, all the drift in the 300 feet melted has been deposited, and chiefly at or beneath the immediate margin of the ice. To the marginal and submarginal accumulations made in this way, the material carried on the ice is added whenever the ice is melted from beneath it. This addition is sometimes considerable and sometimes meagre. If the edge of the ice is without much fluctuation in position, the material dumped over its end may take the form of a narrow ridge or bowlder-wall (*Geschiebe-wall*). If a glacier pushes material in front of it, this, too, becomes a part of the general terminal aggregation of drift.

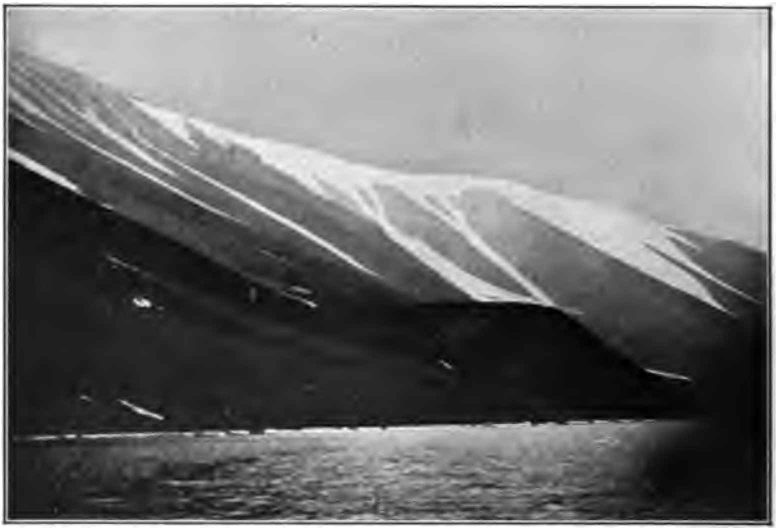
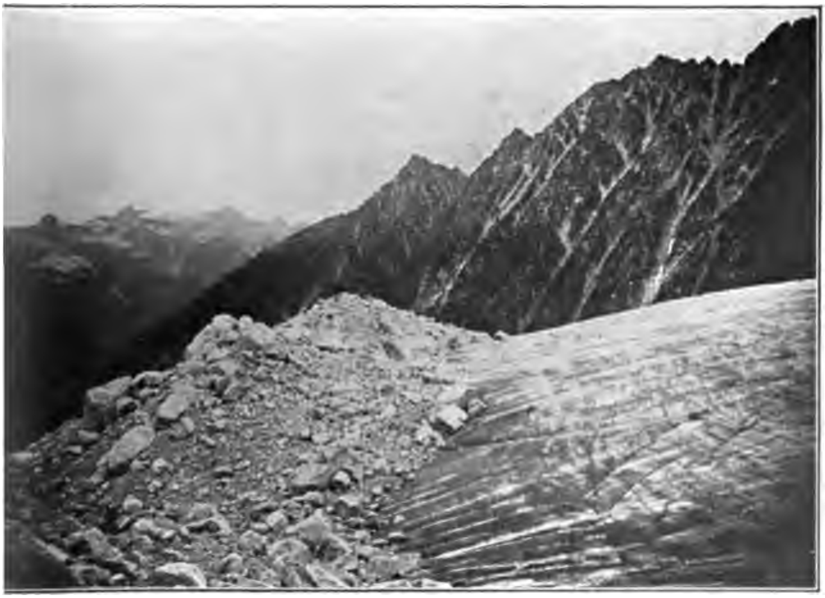


FIG. 276.—Embankment completed. Near the last.

TYPES OF MORAINES.

The terminal moraine.—The thick accumulation of drift made at e end of a glacier or at the edge of an ice sheet, especially where its d or edge is stationary, or nearly stationary, for a considerable time, the *terminal moraine.* That part of the aggregation deposited beneath e ice is sometimes called the *lodge* moraine (Figs. 275 and 276; see also ig. 235); that carried on the ice and dropped at its edge, the *dump* oraine; and that pushed before the ice, the *push* moraine. Many oraines marginal to the ice appear to be push moraines, when they e really lodge moraines from which the ice has withdrawn (Fig. 277).



277.—End of a glacier a few miles west of Kaslo on Lake Kootenai, B. C. A odge moraine from which the ice has withdrawn, giving it the appearance of a ush moraine. It is possible that the lodge moraine material has been pushed p a little by re-advance of the ice. (Atwood.)

push moraine can rarely be distinguished, and the dump moraine o means always, after the disappearance of the ice.

The ground moraine.—When a glacier disappears by melting, all its ris is deposited. All the drift deposited beneath the advancing ice and all deposited from the base of the ice during its dissolution constitutes the *ground moraine.* The thickness of the ground moraine is notably unequal. In general, it is thicker toward the terminus of the glacier and thinner toward its source, but considerable portions of a glacier's bed are often left without débris when the ice melts. In general, the terminal moraine is not only thicker, but more irregularly disposed than the ground moraine.

The lateral moraines.—The surface lateral moraines of valley glaciers are let down on the surface beneath when the ice melts out from under them; but the lateral moraines in a valley from which the ice has melted are not merely the lateral moraines which were on the glacier at a given time. They are often far more massive than any which ever existed on the ice itself at any one time (Fig. 278). As a glacier retreats, its lateral moraine material is more or less bunched. Thus if the ice advances 200 feet while its end is being melted back 300 feet, the lateral moraines on the 300 feet melted are concentrated into 100 feet, as they are delivered on to the land by the melting of the ice from beneath. If the retreat of the end of a glacier be very slow, the bunching may be great. But even this cannot explain the massiveness of some lateral moraines. Furthermore, the materials of which many lateral moraines are composed are nearly as well worn as those of the ground moraine. The massive lateral moraines of which this is true are often made up

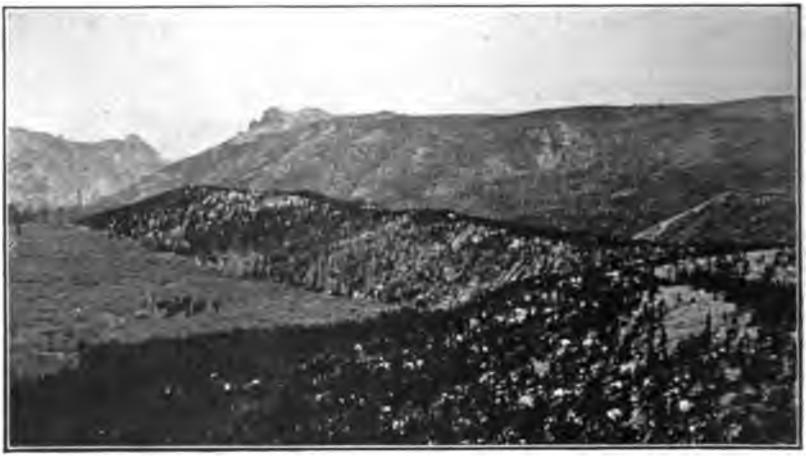


FIG. 278.—A lateral moraine from which the ice has retreated. Bighorn Mountains, Wyo. (Blackwelder.)

chiefly of the drift accumulated beneath the lateral margins of the glaciers. This accumulation is the result of the lateral motion of the ice from center to side. Such sublateral accumulations are akin to terminal moraines. Some of the lateral moraines of ancient valley glaciers, such as those of the Uinta, Wasatch, and Bighorn mountains are several hundred feet high, and in one case about 1000 feet. In northern Italy lateral moraines are said to be 1500 to 2000 feet high.[1]

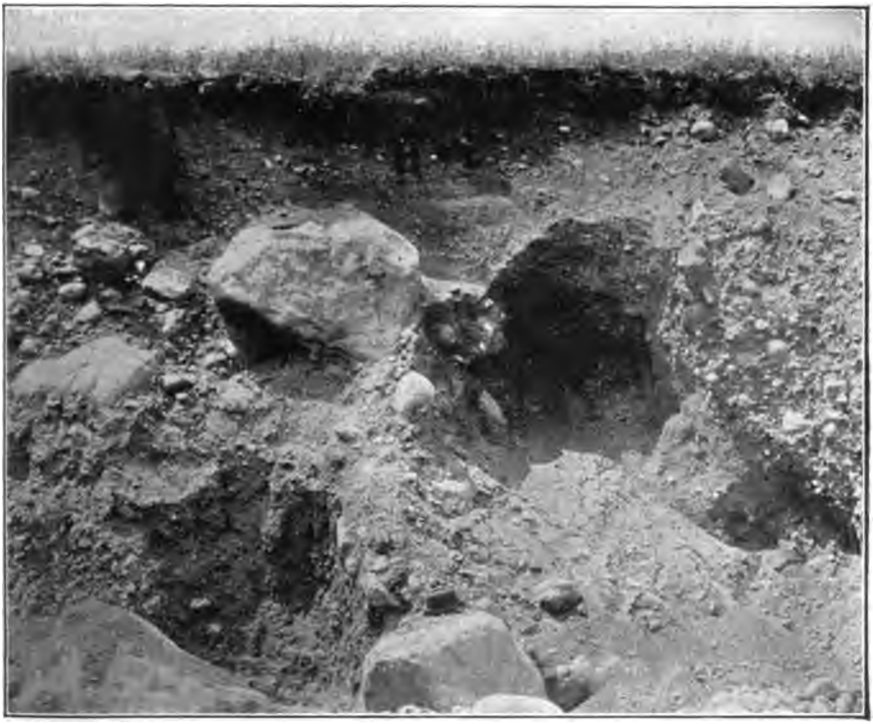


FIG. 279.—Glacial drift, coarse and fine together. (Geol. Surv. of N. J.)

Most of the material which was englacial during the transportation becomes either subglacial or superglacial before deposition, for it ordinarily reaches the bottom or the top of the ice before being deposited. Where the ends or edges of a glacier are vertical or nearly so, as in the high arctic regions, deposition may take place from the englacial position directly.

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[1] Geikie, The Great Ice Age, 3d ed., p. 529.

Distinctive nature of glacial deposits.—The deposits made by glaciers are distinctive. In the first place the ice does not assort its material, and bowlders, cobbles, pebbles, sand, and clay are confusedly commingled (Fig. 279). In this respect, the deposits of ice differ notably from those of water. Furthermore, many stones of the drift show the peculiar type of wear which glaciers inflict. They are not rounded as the stones carried by rivers, though they are notably worn. Many of them have subangular forms with planed and beveled faces, the planes being striated and bruised (Fig. 254). The absence of stratification, the physical heterogeneity, and the striation of at least a part of the stones are among the most distinctive characteristics of glacial drift. A not less real though less obvious characteristic is the constitution of the fine material, for it is in general not the product of rock decay, but of rock grinding. The fine material handled by streams (except glacial streams) on the other hand, is usually the product of rock decay.

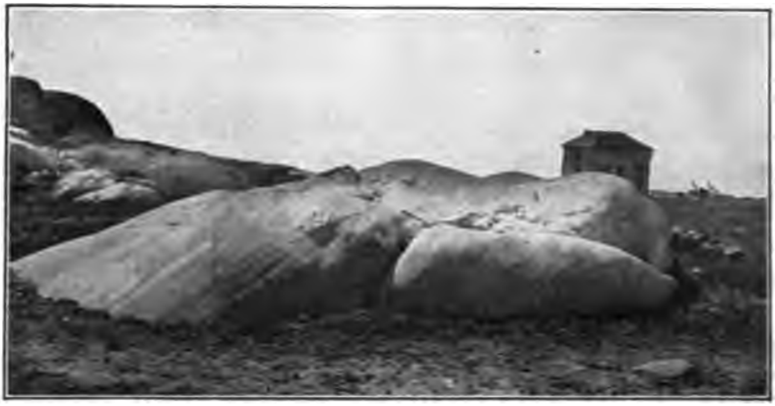

FIG. 280.—Roche moutonnée, Victoria Harbor, B. C.

Glaciated rock surfaces.—Another distinctive mark which a glacier leaves behind it is the character of the surface of the rock on which the drift rests. This is generally smoothed by the severe abrasion to which it has been subjected, and the smoothed surfaces are marked by grooves and striæ, similar to those on the stones of the drift (Figs. 255 and 256). Other distinctive features of a glaciated area are the rounded bosses of rock (*roches moutonnees*, Fig. 280; see also surface about the lakes, Fig. 261), the rock basins, the lakes (Fig. 261), ponds, and marshes, and the peculiar topographies resulting from the unequal erosion, and the still more unequal deposition of the drift. Surface bowlders, often unlike the underlying formations of rock, and sometimes in peculiar and apparently unstable positions, are still another mark of a glaciated area.

GLACIO-FLUVIAL WORK.

The constant but unequal waste of glaciers has already been referred to. The streams to which this gives rise are usually laden with gravel, sand and silt derived from the ice. Since the mud is often light-colored, the streams are sometimes described as "milky." Where the amount of material carried is great, much of it is dropped at a slight distance from the ice, the coarsest being dropped first. Glacial streams are, as a rule, aggrading streams, and therefore develop alluvial plains, called *valley trains* (Fig. 281 and 282), or where they enter lakes (Fig. 283), bays, or other streams, deltas. In its transportation, the river-borne drift is assorted; after its deposition, it is stratified. True glacial deposits in the upper part of a mountain valley are, therefore, often continued below by glacio-fluvial deposits derived from the same source.

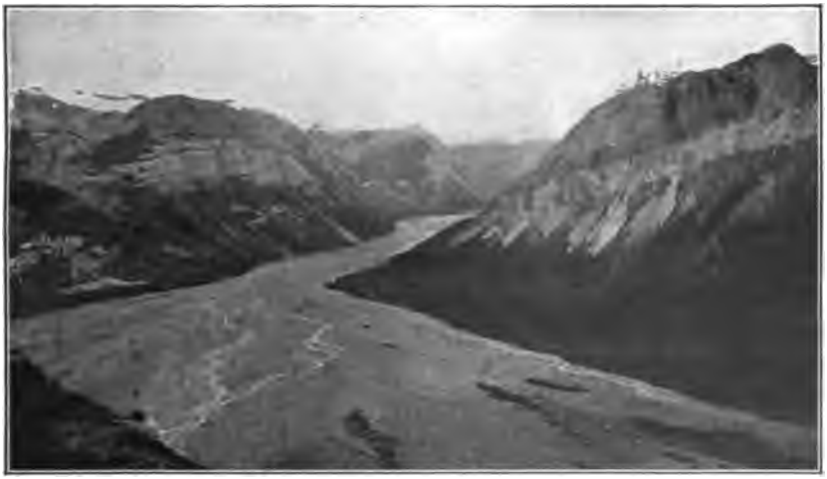

Fig. 281.—Alluviation by glacial stream: Nicolai Creek, Alaska. (Schrader, U. S. Geol. Surv.)

The most common form of such deposit is a *valley train* (Fig. 281) of glacial wash stretching indefinitely down the valley. The silt, sand, and gravel of such trains can usually be distinguished from valley deposits of non-glacial origin by the character of the material, as much of it is the product of grinding, crushing, and fracture, rather than of ordinary surface decay. Its materials are, therefore, fresh and often include rock material which, if long exposed at the surface, would be decomposed or dissolved.

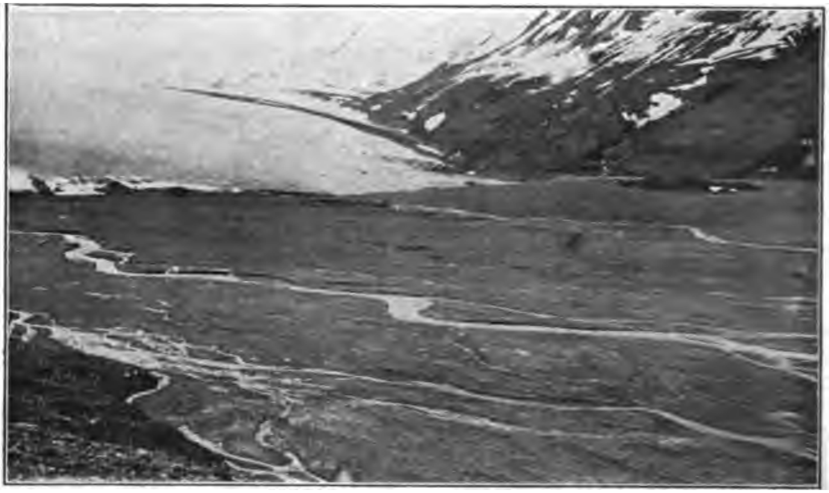

Fig. 282.—Alluviation by glacial stream below Hidden glacier, Alaska. (Gilbert, U. S. Geol. Surv.)

Where an ice sheet ends in a broad face, as did the ancient continental glaciers, numerous streams flow from it and spread their débris in front of the terminal moraine, forming a broad fringing sheet or "apron" (*outwash plain*) along it. Where streams of considerable size form tunnels under or in the ice, these may become more or less filled with wash, and when the ice melts the aggraded channels appear as long ridges of gravel and sand known as *eskers* (*osars* and *serpentine kames* and *kames* of authors. See chapter on glacial period). It has been thought that similar ridges are sometimes formed in valleys cut in the ice from top to bottom, and even that they arise from gravel and sand lodged in superglacial channels. The latter at least is probably rare, as the surface streams usually have high gradients, swift currents, and smooth bottoms, and hence give little opportunity for lodgment.

In the case of ice-sheets, too, in connection with which eskers are chiefly developed, there is usually no surface material except at the immediate edge, where the ice is thin and its layers upturned.

At the mouths of ice-tunnels or ice-channels, especially where they end against terminal moraines, sands and gravels are liable to be bunched in quantity, giving rise, after the adjacent ice has melted, to peculiar hills and hollows of the knob-and-basin type. The hills and short ridges are known as *kames* (see glacial period). Subglacial streams may leave washed and assorted material in their tracks under the ice, and this is sometimes buried under deposits made by the ice itself, so that glacio-fluvial and glacial deposits are interbedded.

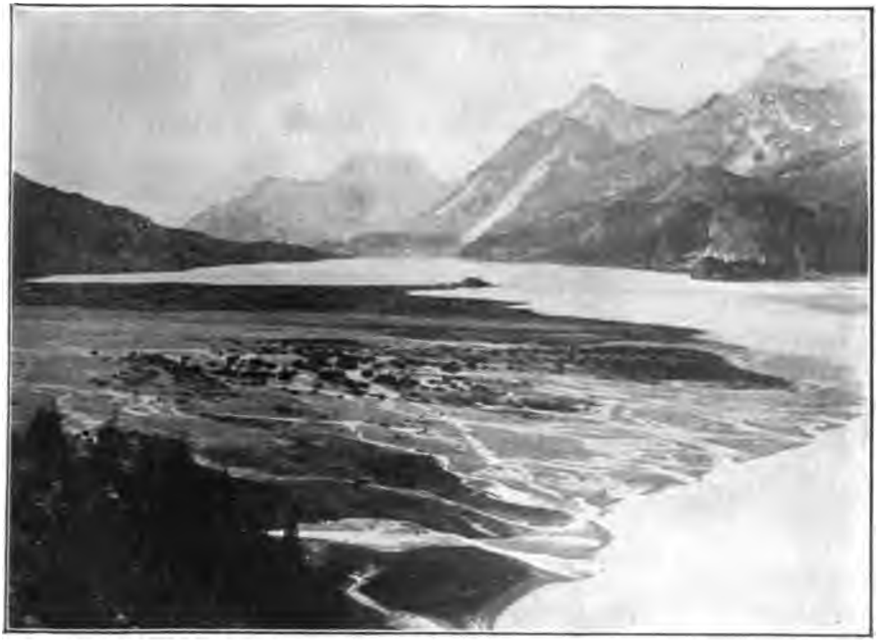

Fig. 283.—Delta at Isola, Lake of Sils, Engadine, Switzerland. (Reid.)

ICEBERGS.

When glaciers advance into water, the depth of which approaches their thickness, their ends are broken off (Fig. 284), and the detached masses float away as icebergs (Fig. 285). Many of the bergs are overturned, or at least tilted, as they set sail. If this does not happen at the outset, it is likely to occur later as the result of the melting and wave-cutting which disturb their equilibrium. The great majority of bergs do not travel far before losing all trace of stony and earthy débris, but the finding of glacial material in dredgings far south of all glaciers shows that they occasionally carry stones far from land.

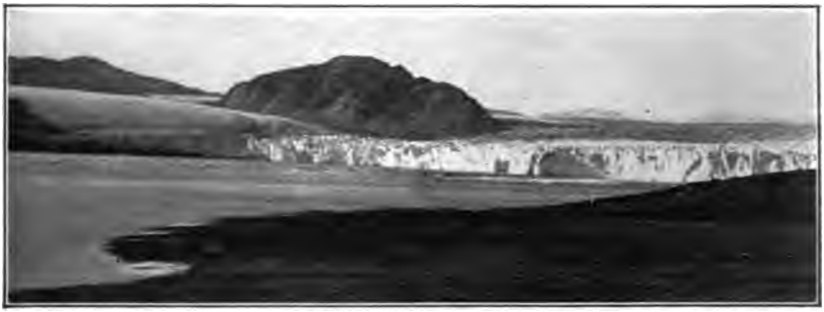

Fig. 284.—End of Muir glacier, Alaska. (Reid.)

THE INTIMATE STRUCTURE AND THE MOVEMENT OF GLACIERS.

With the preceding account of glaciers in mind, we may return to a closer study of their origin, their intimate structure, and their mode of motion. The key to this study is the thesis that a glacier is a mass of crystalline rock—the purest and simplest type of crystalline rock known—since it is made up of a single mineral of simple composition and rare purity, which never appears in a solid state except in the crystalline form.

The growth and constitution of a glacier.—The origin and history of a glacier is little more than the origin and aggregate history of the crystals that compose it. The fundamental conception of a glacier is therefore best obtained by tracing the growth of its constituent crystals. A basal fact ever to be kept in mind is that water in the solid form is always controlled by crystalline forces. When it solidifies from the vapor of the atmosphere it takes the form of separate crystals (Figs. 286-291). Perfect forms are developed only when the flakes fall quietly through a saturated atmosphere which allows them to grow as they descend. Under other conditions, the crystals are imperfect in growth and are mutilated by impact. But however modified, they are always crystals. The molecules are arranged on the hexagonal plan, and, as the expansive power of freezing water shows, the arrangement is controlled by a strong force. Once the definite crystalline arrangement is established, the molecules can be displaced only by relatively great force.

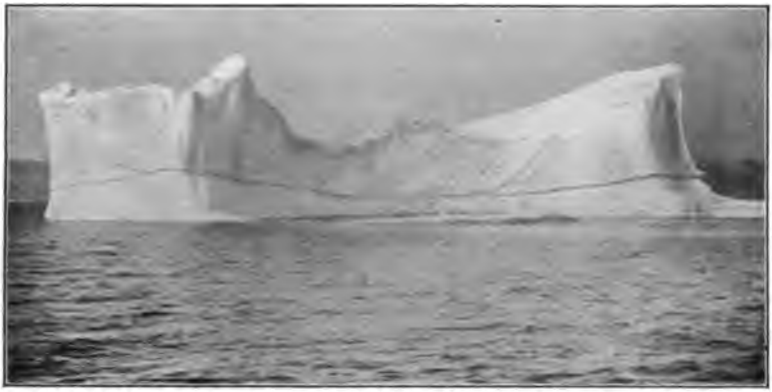

FIG. 285.—An iceberg, west coast of Greenland.

Snow crystals often continue to grow so long as they are in the atmosphere; but if they pass through an under-saturated stratum of air or a stratum whose temperature is above 32° Fahr., they suffer from evaporation or melting. When they reach the ground, the processes of growth and decadence continue, and the crystals grow or diminish according to circumstances.

A glacier is a colossal aggregation of crystals grown from snowflakes to granules of much greater sizes. The microscopic study of new-fallen snow reveals the mode of change from flakes to granules. The slender points and angles of the former yield to melting and evaporation more than the more massive central portions, and this change probably illustrates a law of vital importance. It may often be seen that the water melted from the periphery of a flake gathers about its center, and if the temperature be right, it freezes there. This is a first step toward the pronounced granulation of snow which has lain for some time on the ground. If measured systematically from day to day, the larger granules taken from beneath the surface of this coarse-grained snow are found to be growing. In a series of experiments[1] to determine the law of growth it was found that when the temperature of the atmosphere was above the melting-point the growth was appreciably more rapid than when the air was colder, but there was, on the average, *an increase under all conditions of temperature.* A portion of this average increase of the larger granules appears to come from the diminution and destruction of the smaller ones, for the total number of granules steadily diminishes. A portion of the growth doubtless comes from the moisture of the atmosphere which penetrates the snow and another

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[1] Carried out by C. E. Peet and E. C. Perisho under the direction of one of the authors.

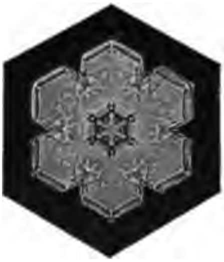
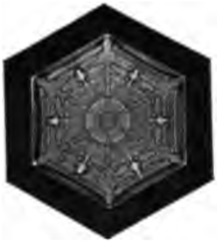


FIG. 286.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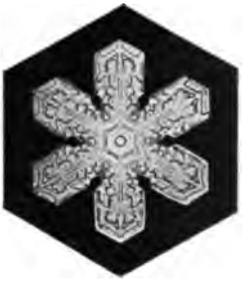
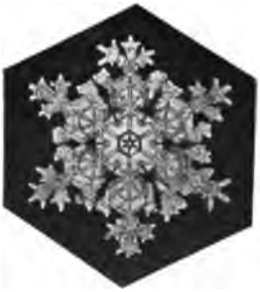


FIG. 288.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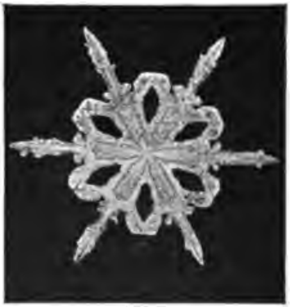
FIG. 290.

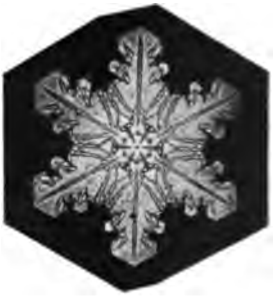
FIG. 291.

FIGS. 286–91.—Snowflakes. (Photographed by W. A. Bentley.)

portion from the moisture derived from surface melting; but beneath the surface of a large body of snow the growth of the large granules is probably chiefly at the expense of the small ones. To follow the process it should be noted that the free surface of every granule is constantly throwing off particles of water-vapor (evaporation); that the rate at which the particles are thrown off is dependent, among other things, on the curvature of the surface, being greater the sharper the curve; that the surfaces of the granules are at the same time liable to receive and retain molecules thrown from other granules, and that, other things being equal, the retention of particles also depends on the curvature of the surface, the less curved surface retaining more than the sharply curved one. Under these laws, it is obvious that the larger granules of smaller curvature will lose less and gain more, on the average, than the smaller granules of greater curvature. It follows that the larger granules will grow at the expense of the smaller. It is also to be noted that, other things being equal, small granules melt more readily than large ones, and that where the temperature is nicely adjusted between melting and freezing, the smaller may lose while the larger gain.

Another factor that enters into the process is that of pressure and tension. The granules are compressed at the points of contact and put under tension at points not in contact, and the pressure and tension are, on the average, likely to be relatively greatest for the smallest granules. Tension increases the tendency to evaporation and adds its effects to curvature, and the capillary spaces adjoining the points of contact probably favor condensation. Ice expands in crystallizing and pressure reduces the melting-point, while tension raises it. The effect of this is slight (p. 276), and it probably plays little part in glacial action, but it is to be correlated with the much more important fact that *compression produces heat* which may raise the temperature of the ice to the melting-point, while tension may reduce the temperature to or below freezing. There is therefore a tendency for the ice to melt at the points of contact and compression, and for the water so produced to refreeze at adjacent points where the surface is under tension. This process becomes effective beneath a considerable body of snow, and here the granules gradually lose the spheroidal form assumed in the early stages of granulation and become irregular polyhedrons interlocked into a more or less solid mass.

A third factor is also to be recognized, though its effectiveness is unknown. Under severe wind pressure, air penetrates porous bodies with appreciable facility. The "breathing" of soils and the curious phenomena of "blowing-wells" and "blowing-caves" teach us of the effective penetration and extrusion of the air under variations of barometric pressure. In the snow-fields, and in the more granular portions of glaciers near their heads, the porosity is doubtless sufficient to allow of the appreciable penetration of the atmosphere. During a part of the time, the probable effect is the condensation within the ice of moisture from the air, and during another part, evaporation from the ice. These alternating processes are attended by oscillations of temperature. While the balance between loss and gain of substance may be immaterial, the oscillating nature of the process and the fluctuations of temperature are probably favorable to granular change.

Whether these processes furnish an adequate explanation of the changes or not, the observed fact is that there are all gradations from snowflakes and pellets into granular névé, and thence into glacier granules (*Gletscherkörner*), varying in size up to that of filberts and walnuts, and even beyond. In coherence, these aggregations may vary from the early slightly coherent granular stage, where the grains are small and spheroidal, to the ice stage, where the cohesion has become strong through the interlocking growths of the large granules. Even when the mass has become seemingly solid ice, sufficient space is usually left between the granules to give the dispersive reflection to light which imparts to glacier ice its distinctive whitish color.

The arrangement of the crystal axes.—The most radical difference between glacier ice and ice formed directly from water is in the arrangement (orientation) of the crystals. In the ice formed on undisturbed water, the bases of the crystals are at the surface and their principal axes are vertical, as shown by Mügge.[1] As they grow, the crystal prisms extend downwards. This gives a columnar or prismatic structure to the ice, well seen when it is "honeycombed" by partial melting. In the glacier, on the other hand, the crystals, starting from snowflakes, have their axes turned in various directions according to the accidents of their fall; and as the snow develops into ice, the principal axes of the

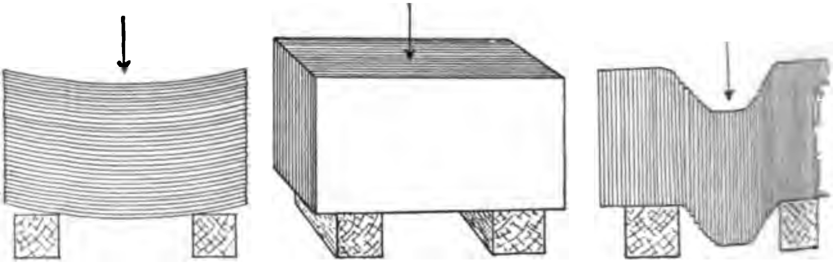

Fig. 291*a*. Fig. 291*b*. Fig. 291*c*.

Figures to illustrate the method of deformation of ice crystals.

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[1] Ueber die Plasticität der Eiskrystalle. Neues Jahrbuch für Mineralogie, etc., 1895, Bd. II, p. 211.

crystals continue to lie in all directions. Hence glacier ice, unlike pond ice, cannot usually be split along definite planes, except where cleavage planes are subsequently developed by extraneous agencies.

While the crystals of a glacier usually have their principal axes in various directions, there appears to be a tendency for them to approach parallelism in certain positions, especially in the basal parts of a glacier near its terminus. Observations on this point are not so full and critical as could be desired, but it is probable that the parallel orientation is partly general, and due to the vertical pressure of the ice, and partly special and local, and connected with the shearing planes and foliation.

The bearing of this partial parallelism of the crystals on shearing and foliation is supposed to reside in the fact that a crystal of ice is made up of a series of plates arranged at right angles to the principal axis of the crystal. These plates may be likened to a pile of cards, the principal axis being represented by a line vertical to them. If a cube be cut from a large crystal of ice, it will behave much like a cube cut from the pile of cards. If the cube be so placed that its plates are horizontal (Fig. 291*a*), and if it be rested on supports at two edges and heavily weighted in the middle, it will sag, the plates sliding slightly over one another so as to give oblique ends, but in this case the cube offers considerable resistance to deformation. If the cube be so placed that the plates stand on edge, each reaching from support to support (Fig. 291*b*), it will offer very great resistance to deformation; but if the plates be vertical and transverse to the line joining the supports, as in Fig. 291*c*, the middle portion will sag under very moderate weighting by the sliding of the plates on one another, and in a comparatively short time the middle portion may be pushed entirely out, dividing the cube. These properties have been demonstrated by McConnel[1] and Mügge, and they appear to throw light on certain phases of the action of glaciers that are most pronounced in their basal parts, and are best illustrated in arctic glaciers.

The Probable Fundamental Element in Glacial Motion.

Melting and freezing.—It has already been shown (p. 279) that the *initial* or *fundamental* cause of glacial motion must be operative at the heads of glaciers where the temperature is lowest and the material most loosely granular. In this condition, there is reason to believe that motion takes place between the grains, rather than by their distortion through the displacement of their laminæ. The fact that the granular structure is not destroyed, as it would be by the indefinite sliding of the crystal plates over each other, sustains this view. The inference is that the gliding planes play a notable rôle in glacial movement only in the basal parts of the lower ends of glaciers, where the greatest thrusts are developed, and where the granules have become largest and most completely interlocked. At the heads of glaciers, where motion is initiated, there may be great downward pressure, but not vigorous thrusts from behind, and probably only moderate thrusts developed within the body itself. There seems therefore no escape from the conclusion that the primal cause of glacial motion is one which may operate even under the relatively low temperatures, the relatively dry conditions, and the relatively granular textures which affect the heads of glaciers. These considerations lead to the view that movement takes place by the minute individual movements of the grains upon one another. While they are in the spheroidal form, as in the névé, this would not seem to be at all difficult. They may rotate and slide over each other as the weight of the snow increases; but as they become interlocked by growth, both rotation and sliding must apparently encounter more resistance. The amount of rotary motion required of an individual granule is, however, surprisingly small, and the meltings and refreezings incident to shifting pressures and tensions, and to the growth of the granules, seem adequate to meet the requirements. In order to account for a movement of three feet per day in a glacier six miles long, the mean motion of the average granule relative to its neighbor would be, roundly, $\frac{1}{10000}$ of its own diameter per day, or one diameter in 10,000 days; in other words, it would *change its relations to its neighbors* to the extent of its diameter in about thirty years. A change of so great slowness under the conditions of granular alteration can scarcely be thought incredible, or even improbable, in spite of the interlocking which the granules may develop. The movement is supposed to be permitted chiefly by the temporary passage of minute portions of the granules into the fluid form at the points of greatest compression, the transfer of the moisture to adjoining points, and its resolidification. The points of greatest compression are obviously just those whose yielding most promotes motion, and a successive yielding of the points that come in succession to oppose motion most (and thus to receive the greatest stresses) permits continuous motion. It is merely necessary to assume that the gravity of the accumulated mass is sufficient to produce the minute temporary liquefaction at the points of greatest stress, the result being accomplished not so much by the lowering of the melting-point as by the development of heat by pressure.

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[1] On the Plasticity of Glaciers and other Ice. Proc. Roy. Soc., Vol. XLIV, 1888, pp. 331–67 (with D. A. Kidd); Vol. XLVIII, 1890, pp. 259, 260; Vol. XLIX, 1891, pp. 323–43.

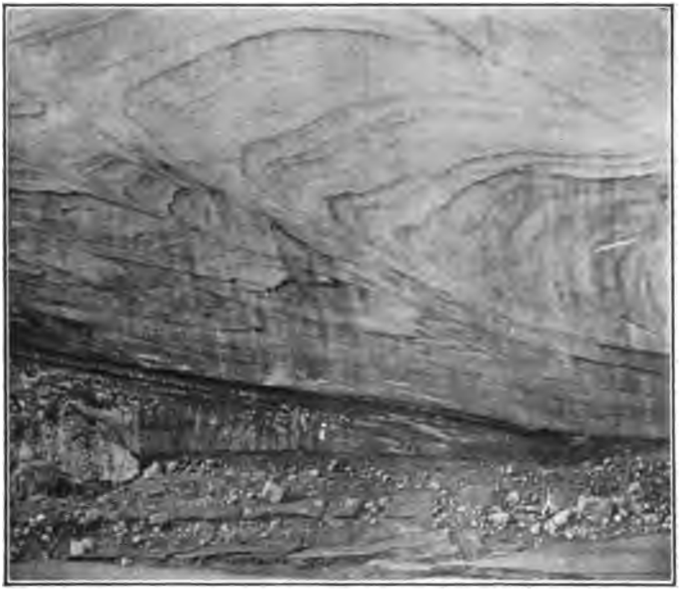

Fig. 292.—Portion of the east face of Bowdoin glacier, North Greenland, showing oblique upward thrust, with shear.

This conception of glacial "flowage" involves only the *momentary liquefaction of minute portions of the mass*, while the ice as a whole remains rigid, as its crystalline nature requires. Instead of assigning a slow viscous fluidity like that of asphalt to the *whole* mass, which seems inconsistent with its crystalline character, it assigns a free fluidity to a succession of particles that form only a minute fraction of the whole at any instant.

This conception is consistent with the retention of the granular condition of the ice, with the heterogeneous (in the main) orientation of the crystals, with the rigidity and brittleness of the ice, and with its strictly crystalline character, a character which a viscous liquid does not possess however much its high viscosity may make it resemble a rigid body.

Accumulated motion in the terminal part of a glacier.—However slight the relative motion of one granule on its neighbor, the granules in any part of a glacier partake in the accumulated motion of all parts nearer the source, and hence all are thrust forward. Herein appears to lie the distinctive nature of glacial movement. Each part of a stream of water feels the hydrostatic pressure of neighboring parts (theoretically equal in *all* directions) and the momentum of motion, but not the rigid thrust of the mass behind. Lava streams are good types of viscous fluids flowing in masses comparable to those of glaciers and on similar slopes, and, in their last stages, at similar rates, but their special modes of flow and their effects on the sides and bottoms of their paths are radically different from those of glaciers. Forceful abrasion, and particularly the rigid holding of imbedded stones while they score and groove the rock beneath, is unknown in lava streams and is scarcely conceivable.

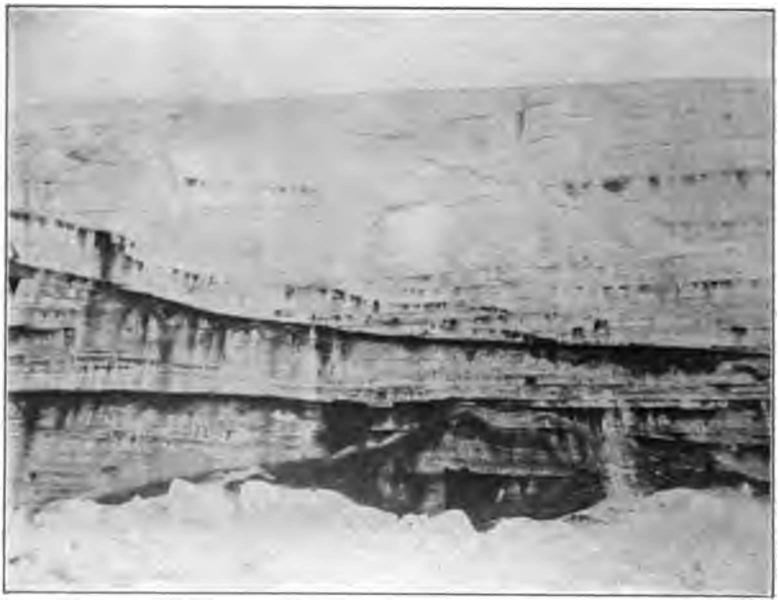

FIG. 293.—Shearing plane well defined. A Spitzbergen glacier. (Hamberg.)

There is, so far as we know, no experimental or natural evidence that any typical viscous body in flowing over a rugose bottom detaches and picks up fragments and holds them as graving tools in its base so fixedly as to cut deep, long, straight grooves in the hard bottom over which it flows. It would seem that competency to do this peculiar class of work, which is distinctive ot glaciers, should be demonstrated before the viscous theory of glacial movement is accepted as even a good working hypothesis. Somewhat in contrast with viscous movement, it is conceived that a glacier is thrust forward rigidly by internal elongation, shears forcibly over its sides and bottoms, and leaves its distinctive marks upon them.

Auxiliary Elements.

Shearing.—In the lower portion of a glacier where normally the thrusts are greatest, the granules fewest, and their interlocking most intimate, shearing takes place within the ice itself. This is illustrated by the accompanying Figs., 292–295. The shearing results in the foliation of the ice and in the forcing of débris between the sheared layers. Thus the ice becomes loaded in a special englacial or baso-englacial fashion, as previously mentioned and illustrated in Fig. 268.

Within the zone of shearing, it is probable that the gliding planes of the crystals come into effective function. It is thought that the combined effect of the vertical pressure, the forward thrust, and the basal drag of the ice, may be to increase the number of granules whose gliding planes are parallel to the glacier's bottom. At any rate, Drygalski reports [1] that there is a tendency to such an arrangement in the basal portion of the Greenland glaciers at their borders. It is conceived that where strong thrusts are brought to bear upon such a mass of granules, those whose gliding planes are parallel to the direction of thrust are strained with sufficient intensity to cause the plates to slide over each other, while those which are not parallel to the direction of thrust are either rotated into parallelism—when they also yield—or are pressed aside out of the plane of shear. As previously noted, shearing is observed to occur chiefly where the ice below the plane of shearing is protected more or less from the force of the thrust. It perhaps also occurs where the basal ice becomes so overloaded with débris that it is incapable of ready movement.

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[1] Grönland-Expedition der Gesellschaft für Erdkunde zu Berlin, 1891–93, Bd. I, p. 491 et seq.

It is also probable that sharp differential strain and shearing are developed at the level where the surface-water of the warm season, descending into the ice, reaches the zone of freezing. The expanding of the freezing water at the upper limit of the cold zone may cause the layer expanded by it to shear over that below. As the level of freezing is lowered with the advance of the warm season, the zone of shearing also sinks. This may be regarded as an auxiliary agency of shearing, of application to a special horizon.

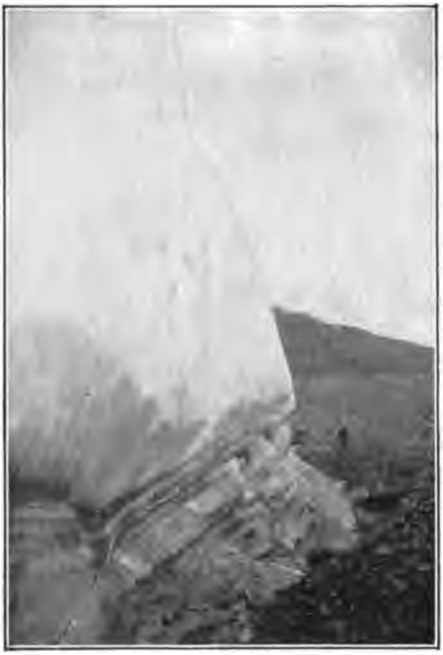

FIG. 294.—Portion of the lateral margin of a North Greenland glacier. Shows upturning of the layers at the base, the cleanness of the ice above the bottom, and, possibly, shearing.

High temperature and water.—In the zone of waste, a higher temperature and more water lend their aid to the fundamental agencies of movement, and there is need for these aids to promote a proportionate movement, for here the granules are more intimately interlocked and the ice more compact and inherently more solid and rigid. The average temperature is, however, near the melting-point (p. 276), and during the warm season the ice is bathed in water so that the necessary changes in the crystals are facilitated, and movement apparently takes place even more readily than in the more open granular portion of lower temperature and dryer state. The extraordinary movements of certain tongues of ice in some of the great fiords of Greenland are probably due to the convergence of very thick slow-moving ice from the interior into basins leading down to the fiords. Into the same basins a large amount of surface-water is concentrated at the same time, with the result that the thick ice, bathed with water and having a high gradient, develops unusual velocity during the warm season.

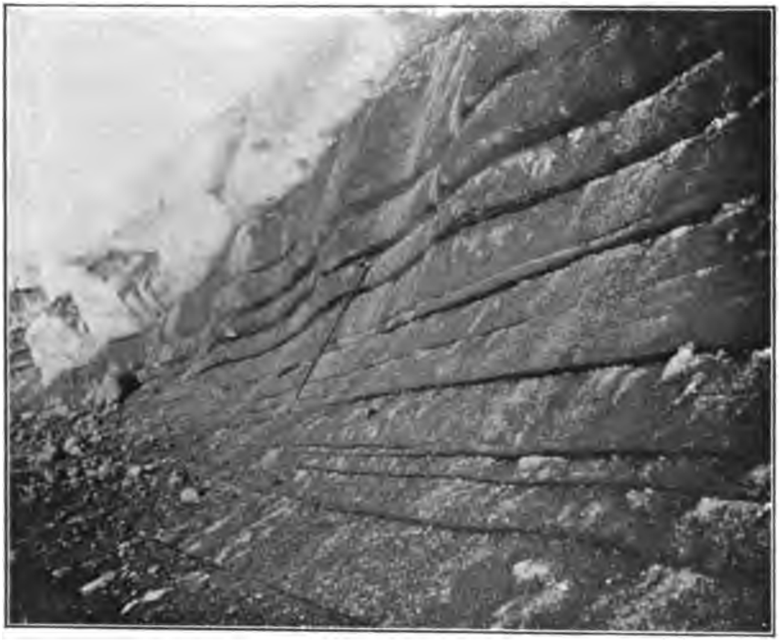

FIG. 295.—Lateral margin of a North Greenland glacier, Inglefield Gulf region The overhanging edges of the successive layers are not altogether the result of shear. They are due in part at least to differential melting along the lines where débris comes to the surface. The débris planes may be shear planes.

Applications.—By a studious consideration of the coöperation of the auxiliary agencies with the fundamental ones, the peculiarities of glacial movement may apparently be explained. In regions of intense cold, where a dry state and low temperature prevail, as in the heart of Greenland, the snow-ice mass may accumulate to extraordinary thicknesses, for the burden of movement seems to be thrown almost wholly upon compression, with the slight aid of molecular changes due to internal evaporation and allied inefficient processes. Since the temperature in the upper part of the ice is very adverse (see p. 277), the compression must be great before it becomes effective in melting the ice, and hence the great thickness of the mass antecedent to much motion. Similar conditions more or less affect the heads of alpine glaciers, though here the high gradients favor motion with lesser thicknesses of ice; but in the lower reaches of alpine glaciers, where the temperatures are near the melting-point, and the ice is bathed in water, movement may take place in ice which is thin and compact.

If the views here presented are correct, there is also, near the end or edge of a glacier, the cooperation of rigid thrust from behind with the tendency of the mass to move on its own account. The latter is controlled by gravity, and conforms in its results to laws of liquid flow. The former is a derived factor, and is a mechanical thrust. This thrust is different from the pressure of the upper part of a liquid stream on the lower part, because it is transmitted through a body whose rigidity is effective, while the latter is transmitted on the hydrostatic principle of equal pressure in all directions.

Corroborative Phenomena.

The conception of the glacier and its movement here presented explains some of the anomalies that otherwise seem paradoxical. While a glacier in a sense flows over a surface, it often cuts long, deep furrows in firm rock. It is difficult to explain this if the ice be so yielding as to flow under its own weight on a surface which is almost flat. If the mass is really viscous, its hold on its imbedded débris should also be viscous, and a bowlder in the bottom should be rotated in the yielding mass when its lower point catches on the rock beneath, instead of being firmly held while a deep groove is cut. This is more to the point since viscous fluids flow by a partially rotary movement. If, on the other hand, the ice is always a rigid body which yields only as its interlocking granules change their form by loss and gain, a rigid hold on the imbedded rock at some times, and a yielding hold at others, is intelligible, for on this view the nature of its hold is dependent on the temperature and dryness of the ice. Stones in the base of a glacier may be held with very great rigidity when the ice is dry, scoring the bottom with much force, while they may be rotated with relative ease when the ice is wet. In short, the relation of the ice to the bowlders in its bottom varies radically according to its dryness and temperature. *A dry glacier is a rigid glacier. A dry glacier is necessarily cold, and a cold glacier is necessarily dry.*

On the view here presented, a glacier should be more rigid in winter than in summer, and the whole thickness of the glacier should experience this rigidity chiefly at the ends and edges, where the relative thinness of the ice permits the low temperature to reach its bottom. The motion in these parts during the winter is, therefore, very small.

In this view may also be found an explanation of the movement of glaciers for considerable distances on upward slopes, even when the *surface* as well as the base is inclined backwards. So far does this go that superglacial streams sometimes run for some distance *backwards*, i.e. toward the heads of the glaciers, while in other places surface-waters are collected into ponds and lakelets. Such a slope of the surface of ice is not difficult to understand if the movement be due to thrust from behind, or if it be occasioned by internal crystalline changes acting upon a rigid body; but it must be regarded as very remarkable if the movement be that of a fluid body, no matter how viscous, for the length of the acclivity is sometimes several times the thickness of the ice. Crevassing and other evidences of brittleness and rigidity find a ready elucidation under the view that the ice is a really solid body at all times, and that its apparent fluency is due to the momentary fluidity of small portions of the mass assumed in succession as compression demands.

In addition to the considerations already adduced, it may be urged that a glacier does not flow as a stiff liquid because its granules are not habitually drawn out into elongated forms, as are cavities in lavas and plastic lumps in viscous bodies. Flowage lines comparable to those in lavas are unknown in glaciers.

All this is strictly consistent with our primary thesis, that a glacier is a crystalline rock of the purest and simplest type, and that it never has other than the crystalline state. This strictly crystalline character is incompatible with viscous liquidity.

Other Views of Glacier Motion.[1]

While these views of glacial motion seem to us to best accord with the known facts, they are not to be regarded as established in scientific opinion, or as the views most commonly held. The mode of glacial motion has long been a mooted question, and is still so regarded. The main alternative interpretations that have been entertained are the following:

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[1] *References on glacier structure and motion.*—L. Agassiz, Études sur les Glaciers, Neuchâtel, 1840. Rendu, Théorie des Glaciers de la Savoie, Soc. Roy. Acad., Savoie, Mém. 1840 (in English, ed. by Geo. Forbes, London, 1874). J. de Charpentier, Essai sur les Glaciers et le terrain erratique du Basin du Rhone, Lausanne, 1841. F. J. Hugi,

(1) In the early days of glacial studies De Saussure thought that glaciers slid bodily on their beds;

(2) Charpentier and Agassiz referred the movement to the expansion of descending water freezing within the glacier;

(3) Rendu and Forbes, followed by many, perhaps most, modern writers, believed ice to be viscous, and that in sufficiently large masses it flows under the influence of its own weight, like pitch or asphalt;

(4) Others, realizing the fundamental difference between crystalline ice and a true viscous body, have fallen back on a vague notion of plasticity which scarcely amounts to a definite hypothesis at all;

(5) Tyndall urged that the movement was accomplished by minute repeated fracturing and regelation, appealing to the fact that broken pieces of ice slightly pressed together at melting temperatures freeze together, but neglecting the fact that this would destroy the integrity of the crystals;

(6) Moseley assigned the movement to a bodily expansion and con-

Ueber das Wesen der Gletscher und Wintereise in dem Eismeer, Stuttgart, 1842. R. Mallet, The Mechanism of Glaciers, Jour. Geol. Soc. Dublin, Vol. I, p. 317; On the Plasticity of Glacier Ice, Jour. Geol. Soc. Dublin, 1845, Vol. III, p. 122; On the Brittleness and Non-plasticity of Glacier Ice, Phil. Mag., XXVI, p. 586. James Thompson, On the Plasticity of Ice as Manifested in Glaciers, Roy. Soc. Proc., Vol. 8, 1857, pp. 455–58. J. Tyndall and T. H. Huxley, On the Structure and Motion of Glaciers, Phil. Trans., 1857, Vol. CXLVII, p. 327. J. D. Forbes, Occasional Papers on the Theory of Glaciers, Edinburgh, 1859. W. Hopkins, On the Theory of the Motion of Glaciers, Phil. Trans., 1862, p. 677; Phil. Mag., 1863, Vol. XXV, p. 224. J. Tyndall, Forms of Water, New York, 1872; The Glaciers of the Alps, London, 1861. James Croll, On the Physical Cause of the Motion of Glaciers, Phil. Mag., 1869, Vol. 38, pp. 201–6. A. Heim, On Glaciers, Phil. Mag., 1871, Vol. 41, pp. 485–508; Handbuch der Gletscherkunde, 1885. H. Moseley, On the Cause of the Descent of Glaciers, Br. Assoc. Rept., 1860, Pt. 2, p. 48; also Phil. Mag., 1869, Vol. 37, pp. 229, 363; Vol. 39, p. 241; Vol. 42, p. 138; Vol. 43, p. 38. Ch. Grad, La Constitution et le movement des Glaciers, Revue Sci., 1872. H. J. Rink, Danish Greenland, 1877. R. M. Deeley, A Theory of Glacial Motion, Phil. Mag., 1888, Vol. 25, pp. 136–64. J. C. McConnel, On the Plasticity of an Ice Crystal, Proc. Roy. Soc. London, Vol. 48, 1890, pp. 256–60; ibid., Vol. 49, 1891, pp. 323–43. O. Mügge, Über die Plasticität der Eiskrystalle, Nachr. k. Ges. d. Wiss., Göttingen, 1895, pp. 1–4. R. M. Deeley and George Fletcher, The Structure of Glacier Ice and its Bearings on Glacier Motion, Geol. Mag. (London), Decade 4, Vol. 2, 1895, pp. 152–62. T. C. Chamberlin, Presidential address before the Geol. Soc. Am., Bull. Geol. Soc. Am., Vol. VI, February 1895, pp. 199–220. Reid, Mechanics of Glaciers, Jour. Geol., Vol. IV, 1896, p. 912. Erich von Drygalski, Grönland-Expedition der Gesellschaft für Erdkunde zu Berlin, 1891–93, Vol. I, 1897.

traction of the glacier, analogous to the creeping of a mass of lead on a roof;

(7) James Thompson demonstrated that pressure lowers the melting-point, and while this effect is so small as probably to be ineffectual, it is correlated with the very important fact that compression may cause melting, which is not the case in most other rocks. He recognized that under pressure partial liquefaction took place, that the water so liberated might be refrozen as it escaped from pressure, and appears to have regarded this as a vital factor;

(8) Croll held that the movement was due to a consecutive series of molecular changes somewhat like the chain of chemical combinations in electrolysis;

(9) Hugi, Eli de Beaumont, Bertin, Forel, and others thought that the growth of the granules was the leading factor in the ice movement;

(10) McConnel and Mügge have made the gliding planes of the ice crystals serve an important function in glacial movement.

It will be seen that the principle of partial liquefaction for which Thompson laid the basis, the crystallization of descending water, urged by Charpentier and Agassiz, and the granular growth on which Hugi, Beaumont, Forel, and others founded their hypotheses, are incorporated in the view already presented. Probably the agencies on which some of the other views are based may also be participants in producing glacial motion, sometimes as incidental factors, and sometimes perhaps as important ones, for under different conditions, different agencies may play rôles of varying importance. For example, in going over the brinks of precipices of sufficient height, glaciers break into fragments which are re-cemented below, and the "reconstructed" glacier moves on as before. Here fracture and regelation are evident. The movement of the gliding planes of the ice crystals over each other, which has been looked upon as a special kind of viscoid movement, probably plays a large part in the shearing movements in certain cases. But neither of these is probably a large factor in ordinary glacial movement, and it seems highly improbable that any of them are essential factors in the primary movements in the snow-fields where glacial action begins.

CHAPTER VI.

THE WORK OF THE OCEAN.

THE general facts concerning the depth of the ocean and the distribution of its water have been given on a preceding page (p. 8), and the origin of the ocean and the ocean basins is discussed in the second volume. This chapter has to do primarily with the processes now going on in the sea and its borders, in so far as they are of importance in the interpretation of geologic history. The study of these processes is prefaced by a few words concerning the amount and composition of the sea-water, the life of the ocean, and the topography of its bed.[1]

Volume and composition.—Every 1000 parts of sea-water contain about 34.40 parts by weight of mineral matter in solution. The principal solids, acids, and bases, combined according to the principles laid down by Dittmar, are shown in the following table:[2]

Chloride of sodium	77.758
Chloride of magnesium	10.878
Sulphate of magnesium	4.737
Sulphate of calcium	3.600
Sulphate of potassium	2.465
Bromide of magnesium	0.217
Carbonate of calcium	0.345
Total salts	100.000

Expressed in terms of tons per cubic mile of sea-water, the composition is as follows:[2]

	Tons per Cubic Mile.
Chloride of sodium ($NaCl$)	117,434,000
Chloride of magnesium ($MgCl_2$)	16,428,000
Sulphate of magnesium ($MgSO_4$)	7,154,000
Sulphate of calcium ($CaSO_4$)	5,437,000
Sulphate of potassium (K_2SO_4)	3,723,000
Bromide of magnesium ($MgBr_2$)	328,000
Carbonate of calcium ($CaCO_3$)	521,000
For sea-water, total dissolved matter	151,025,000

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[1] Much information on these and other points is to be found in the following books: Wild's Thalassa; Thompson's Depths of the Sea; Barker's Deep Sea Soundings, and Maury's Physical Geography; Agassiz' The Three Cruises of the Blake, and the Challenger Reports give much more detailed information concerning these and other matters.

[2] Dittmar, Challenger Reports, Physics and Chemistry, Vol. I, p. 204.

Aside from the ingredients shown in the above tables, the presence of the following has been proved: iodine, fluorine, phosphorus, silicon, boron, silver, lead, copper, zinc, cobalt, nickel, iron, manganese, aluminum, barium, strontium, arsenic, lithium, cæsium, rubidium, and gold. Oxygen, nitrogen, and carbonic acid gas are also present in quantity. The amount of carbonic acid is estimated to be 18 times as great as in the atmosphere.[1]

The amount of sea-water is estimated by Murray at 323,722,150 cubic miles,[2] or about 15 times the volume of the land above sea-level. The volume and composition of the sea-water being known, the amount of mineral matter which it contains may be readily calculated. Assuming the average specific gravity of the mineral matter in solution to be 2.5, the 3.5% by weight becomes 1.4% by volume, and 1.4% of 323,722,150 cubic miles is 4,532,110 cubic miles. This then represents the aggregate volume of mineral matter in the sea if it were precipitated and compacted so as to have an average specific gravity of 2.5. Assuming the average depth of the sea to be 2076 fathoms (12,456 feet), as given by Murray, the mineral matter in solution, if precipitated, would cover the ocean bottom to a depth of about 175 feet. Assuming the area of the land to be to that of the sea as 28 to 72, this amount of mineral matter would make a layer about 450 feet deep over the land. Its amount is equal to about 20% of that of all lands above sea-level, and it falls but little short of that in all lands below 600 feet in altitude. If it were precipitated and concentrated in the shallow waters about the borders of the lands, it would fill the sea out to the depth of about 4000 feet, and would diminish its area by some 19,000,000 to 20,000,000 square miles, an area which is more than ⅓ of the present land surface. In other words, if the mineral matter in the sea-water were precipitated and concentrated in the shallow waters about the lands, it would restore the continental shelves to the land areas, and add an almost equal area beyond.

These comparisons may perhaps help to give some idea of the amount of mineral matter in solution in the sea, but they give no more than a hint of the importance of the solvent power of water in the general processes of rock decay, for most of the substances carried to the sea in solution by rivers are extracted from the water about as rapidly as they are supplied. Thus calcium carbonate is about twenty times as abundant as sodium chloride in river-water,[1] but is only $\frac{1}{225}$ as abundant in sea-water.

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[1] For a discussion of the way in which this gas is held in solution, see Tolman, Jour. of Geol., Vol. VII, pp. 598–618.

[2] Murray, Scot. Geogr. Mag., Vol. IV, p. 39.

The total river discharge into the sea is estimated at 6524 cubic miles of water per year.[2] This water is estimated to carry to the sea annually about half a cubic mile of mineral matter in solution. At this rate it would take about 9,000,000 years for the streams to bring to the sea an amount of mineral matter equal to that it now contains, but the proportions of the ingredients would be very different.

The sodium chloride makes up about 2.4% of the mineral matter in river-water and nearly 78% of the mineral matter of the sea. At this rate it would take nearly 300,000,000 years for the salt of the sea to have been contributed by the rivers. It is not to be understood, however, that this figure indicates the age of the ocean. The salt is not all brought in by the rivers; the rivers have probably not always contributed at the present rate; and much salt once in the sea has been precipitated. Nevertheless the above figure gives some suggestion as to the order of magnitude of the figures which represent the age of the ocean.

In contrast with the salt, the amount of calcium carbonate in the sea is so small that at their present rate of contribution, it would be brought to the sea by rivers in about 62,000 years.

Topography of bed.—The general relations of ocean basins to continents are suggested by Fig. 296. The borders of the continental platforms are covered by the *epicontinental sea*, while the *abysmal sea* occupies the ocean basins proper. From the figure it is seen that an ocean basin is pronouncedly convex upward, and so departs as widely as may be from the current notion of the homely utensil from which it is named. Only when it is remembered that a level surface (on the earth) is one which has the mean curvature of the earth, and that the deeper parts of the ocean basin are well below the mean sphere level, does the current name seem justified.[3] The figure also shows that the depth of an ocean basin is slight compared with the radius of the earth.

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[1] Murray, Scot. Geogr. Mag., Vol. III, p. 76.

[2] Ibid., p. 70.

[3] Limited areas of the ocean bottom are actually concave upward; that is, they are basins in the more commonly accepted sense of the term (see Chapter IX).

The bed of the ocean, like the face of the land, is affected by elevations and depressions, and its deepest points are about as far below its surface as the highest mountains are above it. There are areas of the sea bottom which, as a whole, may be compared to the plains of the land, and others which may be likened to plateaus, and the lines of gradation between them are as ndistinct as they often are on the land. There are mountain peaks, chiefly of volcanic origin, and depressions comparable to the great basins on the land. But apart from these general features, there is little in common between the topography of the sea bottom and that of the land. Mountain systems are, for the most

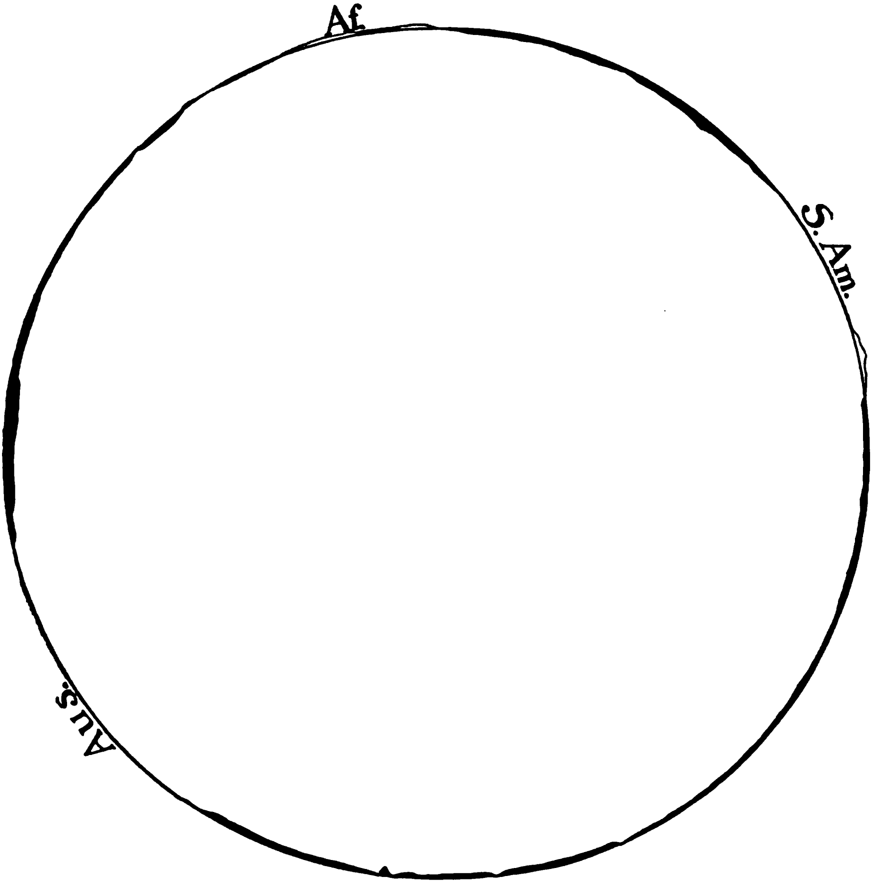

Fig. 296.—General relations of ocean basins to the lithosphere. Lat. 20° S. Depth of the water (black) and height of land exaggerated ten times. (Data from Murray, Scot. Geogr. Mag., Vol. XV, 1899.)

part, absent, though certain islands, like Cuba and some of its associates, may be regarded as the crests of systems which are chiefly submerged. If the water were drawn off from the ocean's bed so that it could be seen as the land is, its most impressive feature would be its monotony. The familiar hills and valleys which, in all their multitudinous forms, give the land surface its most characteristic features are essentially absent. A large part of its surface would be found to be so nearly flat that the eye would not detect its departure from planeness.

The reason for this profound difference is readily found. On the land, the dominant processes which shape the details of the surface are degradational, and though the final result of degradation is flatness (base-level), the immediate result is relief, and, most commonly, relief of the hill-and-valley type. In the sea, the dominant processes are aggradational, and tend to monotonous planeness.

Distribution of marine life.—Marine life has been of such importance in the history of the earth that the elementary facts concerning its distribution and the principles which control it are here recalled. The distribution of marine life is influenced by many factors, chief among which are *temperature* and *depth of water*. Not only is life more abundant in the warmer parts of the ocean than in the colder, but the species inhabiting cold waters are different from those in warm, and few species range through great variations. Many forms of life are restricted to shallow water. Many more, especially those which do not live on the bottom, swim about freely without reference to the depth of the water beneath them, while relatively few are restricted to great depths. Many species are also influenced by the *salinity of the water*, which varies notably along coasts where the fresh waters from the land are discharged; by the *character of the sediment* at the bottom, some species preferring mud, others sand, and others gravel; by the *movement of the waters*, some species preferring still waters and others rough; and some species by the *abundance and nature of the food-supply*, and by *rival and hostile species*.

Subject to the exceptions determined by temperature, etc., plant life abounds in shallow water out to depths of 100 fathoms or so, and is found in abundance at the surface where the depth is much greater. Animal life abounds in shallow water, both at the bottom and above it, out to depths of 200 or 300 fathoms, and occurs in great profusion in the surface-waters of temperate and tropical regions without regard to the depth. The great body of the ocean water lying below a depth of some few hundred fathoms is nearly tenantless, though life reappears sparingly at the bottom, even where the depth is great. For further discussion of this topic, see Chapter XI.

PROCESSES IN OPERATION IN THE SEA.

Within the area of the sea, as on the area of the land, three sets of processes are at work—*diastrophism*, *vulcanism*, and *gradation*.

Diastrophism (p. 2) affects the sea-bottom as the land, but the results are notably different in certain respects. So far as the lithosphere is concerned, the sea-level may be said to be the critical level. At and above it, many processes are in operation which do not appear below, and below it, many which do not take place above. Changes of level which do not involve the submergence of areas which were land, or the emergence of areas which were under water, are relatively unimportant, compared with those which effect such changes. The rise of the bottom of the sea from a depth of 500 fathoms to a depth of 200 fathoms would not lead to important consequences, so far as the area itself is concerned, while an equal rise of the bottom beneath 200 fathoms of water, or an equal subsidence of land 500 feet high, would be attended by more striking consequences. It follows that the changes effected by diastrophism are much more obvious along coasts than in the deep seas. Emergence or submergence shifts the zones of aggradation and degradation, shifts the zone of contact of ocean and land, and changes the region concerned from one appropriate for sea life to one appropriate for terrestrial forms, or *vice versa*.

Over the continental shelves the water is shallow and the bottom relatively smooth. If a coastal region be elevated evenly, or if the sea-level be drawn down, the new shore-line on the smooth surface of the former submerged shelf will be relatively regular, even though the coast was notably irregular before the change. Thus in Fig. 297 the coast-line is notably irregular. A sea-withdrawal or a land-uplift of 120 feet would change the coast-line to the position of the 20-fathom line, when it would be notably less irregular than now. If it were shifted to the 100-fathom line, few irregularities would remain. In so far as new coast-lines formed by the lowering of the sea (or rise of the crust) depart from straightness, it is usually by broad, smooth curves. Local

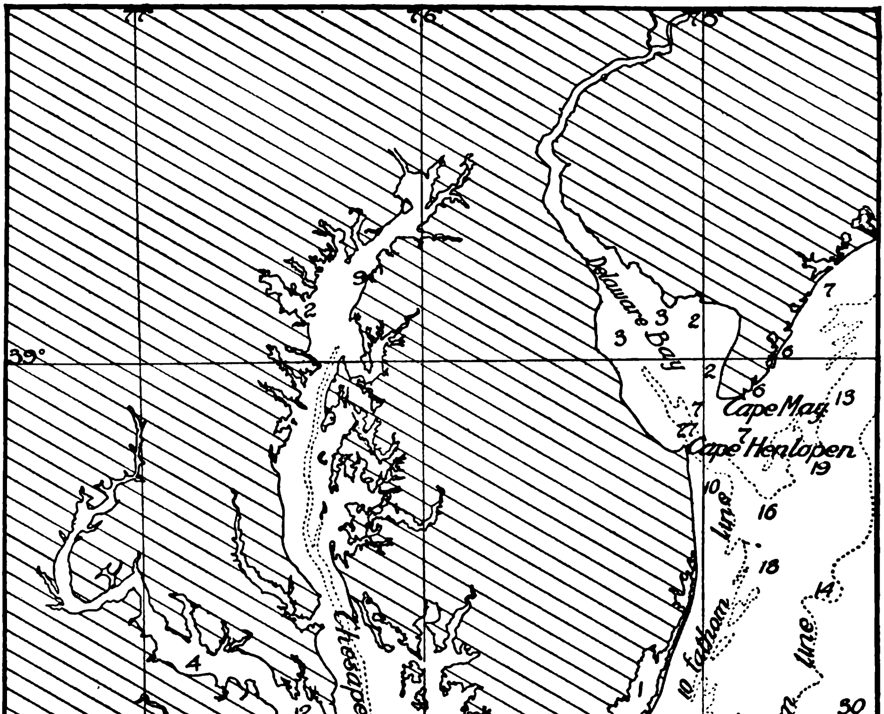

Fig. 297.—Sketch of the eastern coast of the United States from Cape May to Cape Henry, showing coastal irregularities. The figures represent the depths of water in fathoms. (From charts of C. and G. Surv.)

uplifts of coastal lands, and especially uplifts along axes normal to the trend of the coast, would give rise to projections of land, and so to coastal irregularities; but such uplifts are rarely so localized as to give origin to minor projections. It follows that rising coasts, and those which have recently risen, or more likely, coasts along which the sea-level is sinking or has recently sunk, are likely to be regular so far as details of outline are concerned. Subsidence of a coast-line (or rise of the sea-level) tends to the opposite results, for in this case the sea advances on a surface which has more or less relief, and the water takes possession of every depression brought to its level. The lower parts of the valleys are converted into bays, the length and width of which depend on the slope and width of the valleys drowned. The numerous bays at the debouchures of the streams along the Atlantic coast of the United States, from Long Island Sound to Carolina, such as the Delaware, Chesapeake, (Fig. 297) and numerous smaller bays, are the results of recent sinking, which has allowed the sea to invade the lower ends of river valleys.

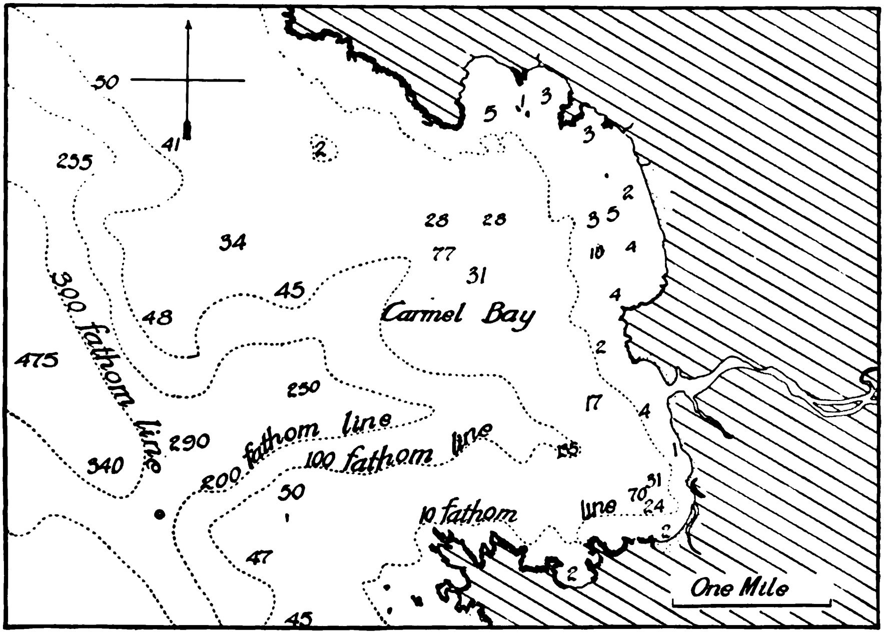

FIG. 298.—Sketch of Carmel Bay, Cal. The contours below sea-level show a deep submerged channel. (From charts of C. and G. Surv.)

The ragged coast of Maine is another example, though glaciation as well as subsidence has been operative here. From the present configuration of coast-lines, it has been inferred that the present is, on the whole, an era of continental depression.[1] River valleys, the lower ends of which are embayed, are sometimes found to be continuous with submerged valleys beyond the coast-line (Fig. 298). Submerged river valleys show that the surface in which they lie was once land.

Bays may be developed by local subsidence as well as by the submerging of valleys, though decisive examples are not readily cited. Bays may also be produced by uplift of the surface on either side of an area which does not change its level. For example, uplift on either side of the Gulf of California has probably been one element, though probably not the only one, in the development of this indentation. The general outline of a great bay produced by coastal warping might be regular, though it would be likely to be marked by small irregularities where the streams enter. It is not to be understood that all, or even most, bays are due to local diastrophism.

Diastrophism, then, as it affects the ocean-bottoms and the ocean-borders, may make the water of any ocean shallower or deeper; it may cause the emergence or submergence of land; it may make coast-lines regular or irregular; it may shift the habitat of life, and through these changes may greatly influence the processes of gradation, which are especially active along the contact of sea and land.

Vulcanism affects the sea-bottom much as it affects the land. At the volcanic centers, where the great body of extruded matter accumulates, mounds and mountains are built up. Most of the mountain peaks of the sea-bottom, whether their crests are islands, or whether they are wholly submerged, have had a volcanic origin. The rock material ejected from submarine vents is probably less widely distributed than that from vents on land, and so far forth, the volcanic cones in the oceans are steeper than those on land. Where volcanic cones are built up near the surface of the sea, they often furnish a home for shallow-water life, such as polyps. Wherever built up so as to be within the reach of waves, gradational processes are stimulated.

The processes of vulcanism do not commonly influence coasts of continents directly, for few volcanoes lie immediately on coasts. In places, however, as at various points in and about Italy, the configuration of the coast is influenced by the building of volcanoes. Indirectly, vulcanism influences the shape of coast-lines, for the resistance of igneous rock is often different from that of the rock with which it is associated, and under the influences of the forces of gradation it may come to form projecting points or reentrants, as the case may be.

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[1] J. Geikie. Earth Sculpture, p. 329.

The number of active volcanoes on islands is about 200, or about two-thirds of all now known. Since the area of the sea is about three times that of the land, the known active volcanoes in the sea are rather less numerous per unit area than those on the land. The number of active vents beneath the sea is altogether unknown. A few submarine eruptions have been observed, and those observed are probably but a small percentage of those which have taken place in historic time. Slight eruptions in deep water might not manifest themselves at the surface in an unequivocal way, even were observers stationed near them. Volcanic cones which fail to reach the surface are known, and the forms of many sea-bottom mountain peaks are such as to make it probable that they are volcanic. These phenomena, as well as the numerous volcanic islands, give some indication of the importance of submarine eruptions in past time.

Ocean volcanoes, and especially submarine volcanoes, affect both the temperature and the composition of the sea-water. Both the increase of temperature and the solution of volcanic gases increase the capacity of the water for mineral matter, and both the change in temperature and composition affect the life of the adjacent waters. The destruction of life during eruptions occasions the generation of the products of organic decomposition, and these stimulate further chemical changes. The diffusion of affected waters occasions chemical changes wherever they go. The effects of oceanic volcanoes on the sea-water are, therefore, appreciable, when long periods of time are considered. The deposition of the finer parts of volcanic discharges will be considered in connection with the deposits of the deep sea.

Gradation.—The gradational processes of the land and the sea are in striking contrast. On the land, degradation predominates, and aggradation is subordinate. In the sea, aggradation predominates, and degradation is subordinate. On the land, degradation is, on the whole, greatest where the land is highest, while aggradation is of consequence only where the land is low, or where steep slopes give place to gentle ones. In the sea, degradation is virtually confined to shallow water, or to what might be called the highlands of the sea, while aggradation is nearly universal, but most considerable in shallow water, or where shallow water gives place to deep. Both the degradational and aggradational work of the sea are greatest near its shores. Opposed as the gradational work of the land and sea are, they yet tend to a common end—the leveling of the surface of the lithosphere.

The gradational processes which affect the sea-bottom may be divided into three categories: (1) Those effected by mechanical means, (2) those effected by chemical means, and (3) those effected by organic agencies.

The mechanical work of gradation in the sea is effected chiefly by the movements of the water, and, very subordinately, by the movements of the ice which the water carries. The results of these movements may be degradational wherever the water is sufficiently shallow for the motion to affect the bottom. Elsewhere it is aggradational.

The direct gradational work effected by chemical means is likewise partly degradational and partly aggradational. If at any time or place the water becomes supersaturated with any mineral substance, precipitation takes place, and the precipitate accumulates as sediment on the bottom. This sometimes happens in lagoons and other small inclosures, and perhaps in open water. On the other hand, wherever solution is effected, degradation is the result. Solution is most important where the bottom consists of relatively soluble rock, such as lime carbonate.

Organic agencies are, on the whole, aggradational. Accumulations of coral, coral débris, shells, etc., help to build up the sea-bottom, and most rapidly in shallow water where the proper forms of life are most abundant. Here also should be mentioned the accumulations of carbonaceous matter, especially in the form of plant bodies. In the aggradation effected directly by organic agencies, the sea is passive. Its only part is to support the life which gives rise to the solid matter, and incidentally to float a part of it in its currents.

MOVEMENTS OF THE SEA-WATER.

The movements of the sea-water fall into several categories. There is (1) a general circulation of sea-water, determined chiefly by three factors: differences in density in the sea-water, differences of level, and the general movements of the atmosphere; (2) periodic movements which are not primarily circulatory, brought about by the attraction of the sun and moon; and (3) aperiodic movements, due to occasional causes, such as earthquakes, volcanic explosions, landslides, etc., which determine local and temporary movements, often of exceptional strength.

Differences in density and their results.—Differences in density result from differences in temperature and salinity. Temperature alone considered, water would be densest where it is coldest, namely in the polar regions. Differences in salinity result from differences in evaporation and from inequalities in the supply of fresh water. Evaporation alone considered, the sea-water should be densest where evaporation is greatest; but the equatorial region, where evaporation is greatest, is also a region where precipitation is heavy, and precipitation, by freshening the water, opposes the effect of great evaporation. The greatest differences in density due to the unequal supply of fresh water are to be found near the borders of continents, where the precipitation on the land is discharged into the sea. In the polar regions, the great supply of fresh water, especially during the season when the ice is melting, opposes the effect of the low temperature, so far as the density of the water is concerned. The result of the operation of these factors affecting the density of the sea-water is to insure a general circulation, directed to the end of equalizing the densities; and since the disturbing factors are constantly in operation, equilibrium is never established, and the movements of the water are perpetual.

The pressure gradients resulting from differences of density are so slight that the resulting movements are scarcely more than a creep of the waters. In general they are far too slow to be of importance in gradational work; but the earth's rotation deflects the creeping waters and tends to concentrate the equator-ward movement into currents on the east sides of the continents, and the pole-ward movement on the west sides. In favorable situations these currents may be competent to produce sensible mechanical results. Even where this is not the case the circulation helps to equalize the temperatures of the sea, and so of the air above and of the land about. Indirectly, therefore, the circulation of the ocean-waters affects every geological process which is sensitive to climate.

Differences in level and their results.—While the surface of the ocean is the common datum plane to which elevations and depressions are referred, it is to be remembered that the sea has "a very complicated undulating surface in consequence of the attraction which the heterogeneous and elevated portions of the lithosphere exercise on the liquid hydrosphere. In the opinion of geodesists, the geoid may in some places depart from the figure of the spheroid by 1000 feet."[1] These variations in level would, however, not occasion circulation. The differences in level which determine circulation are much more trivial. Every stream which pours fresh water into the sea tends to raise the level of the water where it enters. The waters brought to the ocean by the Amazon, the Mississippi, and other great rivers would appreciably change the level of the sea at their debouchures, if the excess did not promptly flow away. The ready mobility of the water, however, prevents its accumulation, and the discharge of every stream generates widespread movement. This movement is strongest at the debouchure, and weakens with increasing distance from it, though in the case of great streams, such as the Amazon, the movement is traceable, by means of the sediment which the water carries, hundreds of miles out to sea.

Changes of level are also brought about by the winds, which pile up water along the shore against which they blow. The level of the water is said to have risen 24 feet at Calcutta on October 5, 1864, as the result of a severe storm. While this is exceptional, a rise of 2 feet is not rare. This piling up of the waters along shore insures a compensating movement (undertow, littoral currents, etc.) in some other direction. Unequal evaporation and precipitation likewise disturb the level of the sea and occasion movement. In the open sea the movements generated by differences of level, like those generated by differences of density, are chiefly slow, creeping movements, but movements which never cease. In bays and gulfs, on the other hand, the surface of the water may be so raised, either as the result of wind, river discharge, or heavy precipitation, as to give rise to strong outward currents. There is little doubt at the present time that the Gulf Stream owes its origin primarily to the difference of level between the Gulf of Mexico and the Atlantic.[2]

Movements generated by winds.—The circulation resulting from the tendency of the winds to change the level of the sea-water has already been mentioned, but the wind also works in other ways. Where the winds have a somewhat constant direction and are at the same time strong, they determine a general movement of the surface-waters in their own direction, the surface-water being dragged along at a rate somewhat less than that of the wind itself. The constant trades appear to be the chief generators of the equatorial ocean-currents. Once generated, these currents may be concentrated and their courses modified. The currents generated by trades are turned north and south when directed against a continent; they are modified by the configuration of the bottom if the water be shallow, and always and everywhere, except at the equator, they are deflected by the rotation of the earth, in the northern hemisphere to the right, and in the southern to the left. The pole-ward currents generated in the equatorial region by the trades, and directed by the winds, the lands, the configuration of the bottom, and the rotation of the earth, determine compensating currents from high latitudes to low, and the same influences which control the course of the former direct the latter as well.

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[1] Murray. Scottish Geographical Magazine, Vol. XV, p. 507.

[2] Lindenkohl. Science, Vol. X, 1899, p. 807.

Since the atmospheric movements are so far constant that there is a prevailing direction of winds in all latitudes, the winds, as well as differences of density and differences of level, insure a general and continual circulation of sea-water. The geological effects of this circulation are direct and indirect; direct, by gradation of the bottom over which they flow, and indirect, by the modifications of climate they produce. Since rotation deflects the pole-ward currents to the east sides of the oceans (west sides of the continents) and the equator-ward movements to the west sides of the oceans (east sides of the continents), the east shores of the oceans are warmer than the west in corresponding latitudes, and the west sides of the continents are both warmer and moister[1] than the east sides.

The most obvious disturbance of sea-water resulting from the winds is the generation of *waves*. Waves are not primarily parts of the general oceanic circulation. Since they are generated in other ways than by winds, and since the gradational effects of waves are independent of their origin, the effects of wind-waves will not be considered separately.

Movements generated by attraction.—One of the movements of the sea-water which is not primarily circulatory results from the attraction of the moon and sun. The tide is really the result of the *inequalities* of the attraction of these bodies on different parts of the earth. The lunar tide is more important than the solar, not because the attraction of the moon is greater, for it is not, but because its differential attraction, the result of its lesser distance, is greater.

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[1] This does not hold for tropical latitudes.

The distance of the moon from the earth is about 240,000 miles. If this be taken as the distance from the center of the moon to the center of the earth, 236,000 and 244,000 miles respectively are the distances from the center of the moon to the nearest and most distant points on the earth. The distance of the sun from the earth is about 93,000,000 miles. If this be taken as the distance between the centers of these bodies, then the distances from the center of the sun to the nearest and most distant points on the earth's surface are 92,996,000 and 93,004,000 miles respectively. The ratio of 4000 to 236,000 or to 244,000 is much greater than the ratio of 4000 to 92,996,000 or to 93,004,000. Hence the tide-producing force of the moon is greater than that of the sun.

The tides show themselves along shores in the form of waves which, in shallow water, become translatory. They differ from the wind-waves in their periodicity, and locally in their greater height. The effects of the tidal waves on the shores of the sea, and on the bottom in shallow water, are the same as the effect of wind-waves of equal strength, and need not be separately considered in connection with the gradation of the sea-bottom. In passing through narrow straits or narrow passes of any sort, the tidal movement becomes a current which, under favorable conditions, abrades or "scours" the bottom effectively. The tidal currents in the narrow passes about New York harbor may serve as an illustration.

Aperiodic movements.—In addition to the wind-waves which are essentially constant and universal, and to the tidal waves, which are periodic, there are accidental waves which are locally and temporarily of importance. Such are earthquake-waves, which are sometimes extremely destructive. Thus an earthquake-wave on the coast of Peru in 1746 swept a frigate several miles inland and deluged Lima, seven miles from the shore. The havoc of most earthquakes affecting coasts, such as that of Libson in 1755, is greatly aggravated by accompanying sea-waves. Earthquake-waves differ from ordinary waves in being translatory, and so in being more effective on the bottom in deep water. Their greatest force, however, is felt in shallow water and on shores. Volcanic eruptions likewise give rise to exceptional aperiodic waves.

The same is true of landslides where they affect the coast or any part of the sea-bottom. The fall of glacier ends and the capsizing of icebergs likewise generate strong waves. To the category of exceptional waves also belong those generated by the winds of exceptional storms, such as that which devastated Galveston in 1900.[1]

Summary.—From the point of view of their direct geological results in shallow water, all movements of the sea-water may be grouped into two main classes—(1) waves, with the undertow and the littoral currents they generate, and (2) ocean-currents.[2]

WAVES.

Wave-motion.[3]—The most common waves, and from the present point of view the most important, are those generated by winds. During the passage of a wave, each particle affected by it rises and falls, and moves forward and backward describing an orbit in a vertical plane. If the passing wave is a swell, the orbit of the particle is closed and is either a circle or an ellipse; but in the case of a wind-wave the orbit is not closed. In such a wave two things move forward, the undulation and the water. The velocity of the undulation is relatively rapid; that of the water, slow and rhythmic. On the crest of the wind-wave each particle of water moves forward, and in the trough it moves less rapidly backward, and the excess of the forward movement over the backward gives it a slight residual advance. This residual advance is the initiatory element of current. By virtue of it, the upper layer of water is carried forward with reference to the layer below, in the direction toward which the wind blows. The waves of any considerable or long-continued wind, therefore, generate a current tending in the same direction as the wind.

The agitation of which waves are the superficial manifestation is not restricted to the surface, but is propagated indefinitely downward. Near the surface the amount of motion diminishes rapidly with increasing depth (Fig. 299), but the rate of diminution itself diminishes, and

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[1] National Geographic Magazine, Vol. XI, pp. 377–392.

[2] For causes of ocean-currents, see Croll's Climate and Time: Proc. Roy. Soc., 1869–73, and Jour. Roy. Geog. Soc., 1871–77.

[3] In the following pages concerning the waves and their work Gilbert's classic discussion of shore features, in the Fifth Annual Report of the U. S. Geol. Survey, pp. 80–100, is freely drawn on. Another incisive discussion of certain shore phenomena is that of Fenneman, Jour. of Geol., Vol. X, pp. 1–32.

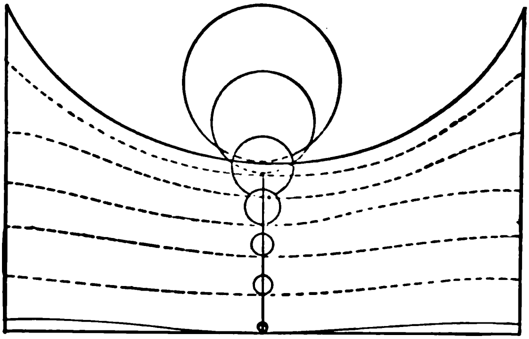

FIG. 299.—Figure illustrating the decrease in the amount of wave-movement with increase of depth. (Fenneman.)

there seems no theoretic reason for assigning any definite limit to the downward propagation of the oscillation.

At the surface, the radius of the circular orbit which a particle of water in a wave tends to describe is half the height of the wave. At a depth equal to one wave-length, the radius of the circle described by a particle is $\frac{1}{535}$ as great as at the surface, and at a depth equal to two wave-lengths, $\frac{1}{300000}$. If the height of a wave be 43 feet, the radius of the circle described by a surface particle is $21\frac{1}{2}$ feet. If the length of the wave be 300 feet, the radius of a particle at a depth of 300 feet is only about $\frac{4}{10}$ of an inch, and at 600 feet $\frac{1}{1200}$ of an inch.[1] These figures make it clear that effective agitation of the water does not extend to great depths.

So long as the velocity of the wind remains constant, the velocity of the current which the wind-waves generate is less than that of the wind, and there is always a differential movement of the water, each layer moving faster than the one beneath. The friction is thus distributed through the whole vertical column of the water in movement, and is even borne in part by the sea-bottom if the movement extends so far down. The greater the depth, the smaller the share of the friction each layer of water is called upon to bear, and the greater the velocity of the current generated by a given wind. But while the wave-motion extends indefinitely downward, the lower limit of agitation effective in erosion is soon reached. Engineering operations have shown that submarine structures are little disturbed at depths of five meters in the Mediterranean and eight meters in the Atlantic.[1] On the other hand, débris as coarse as gravel, which is transported by rolling on the bottom, is not infrequently carried out to depths of 50 feet, and sometimes even to 150 feet. Fine sediment, like silt, is disturbed at still greater depths, for ripple-marks, which indicate agitation of the water, are said to have been found at depths of 100 fathoms.[2]

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[1] Dana. Manual of Geology, 4th ed., p. 213.

When a wave approaches a shelving shore, its habit is changed. The velocity of the undulation is diminished, while the velocity of the advancing particle of water in the crest is increased; the wave-length, measured from trough to trough, is diminished, and the wave-height is increased; the crest becomes acute, with the front steeper than the back, and these changes culminate in the breaking of the crest, when the undulation proper ceases. Waves of a given height break in about the same depth of water, and the line along which incoming waves break is the line of *breakers*. The line of breakers is in deeper water and farther from shore when the waves are strong than when they are weak. Waves are reported to have broken in 100 fathoms of water,[3] but this must be regarded as very exceptional. The return of the water thrown forward in the crests of waves is accomplished by a current along the bottom called the *undertow*. The undertow is sensibly normal to the coast when uninfluenced by oblique waves, and is efficient in removing the products of erosion.

Since the incoming wave affects water which is at the same time under the influence of the undertow, it gives to that current a pulsating character, for the wave-motion sometimes supports and sometimes opposes the undertow, and thus endows it with a higher transporting power than belongs to its mean velocity. Near the breaker-line, the oscillations communicated by the wave may momentarily overcome and even reverse the movement of the undertow. Inside the breaker-line, irregular oscillation only is communicated. The broken wave-crest, dashing forward, overcomes the undertow and throws it back, and the water returns as a simple current descending a slope. The power of the undertow diminishes rapidly from the breaker-line outward as the depth of the water increases.

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[1] Delesse. Lithologie des Mers de France. Cited by Geikie, Text-book of Geology, 3d ed., p. 438.

[2] Sir G. Airy. Encyclopedia Metropolitana, Art. Waves. Cited by Geikie, loc. cit., p. 438.

[3] Stevenson. Treatise on Harbors.

When waves advance on the shore obliquely, a shore-current is developed as illustrated by Fig. 300, where *ab* represents the direction of the incoming wave, *bc* the direction of the littoral current, and *bd* the direction of the undertow. Where they strike the borders of land, the wind-waves, therefore, generate two other movements, the undertow and the littoral current. Any particle of water near shore may be affected by any two or by all three of these movements at the same moment. The effect of littoral current and undertow is to give a particle of water on which both are working a direction between the two, as *be*. The effect of other combinations can be readily inferred. These various combinations are of consequence in the transportation of débris.

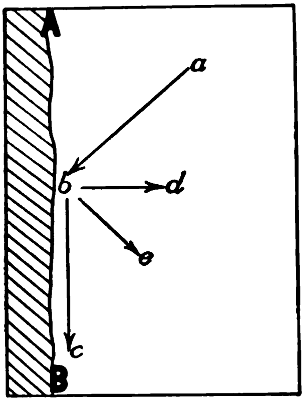

Fig. 300.—Diagram showing relative directions of wave, undertow, and shore-current.

WORK OF THE WAVES.

Erosion.

The general effects of the waves and the other movements to which they give rise along shores are (1) the wear of the shores; (2) the transportation for greater or less distances of the products of wear; and (3) the deposition of the transported materials.

By waves and undertow.—In the dash of the waves against the shore, the chief wear is effected by the impact of the water and of the débris which the water carries. Lesser results are accomplished in other ways.

When the land at the margin of the water consists of unconsolidated material, or of fragmental material but slightly cemented, the impact of the water is sufficient to displace or erode it. If weak rock be associated with resistant rock within the zone of wave-work, the removal of the former may lead to the disruption and fall of the latter, especially n weak rock is washed out from beneath the strong. The impact of the water is competent also to break up and remove rock which was once resistant, but which has been superficially weakened by changes of temperature. Rock affected by numerous open joints is likewise attacked with success, for by the dash of the waves the blocks between the joints may be loosened and literally quarried out. It may, however, be doubted whether the dash of waves of clear water, even when their force is many tons to the square foot, has any appreciable power to wear rock which is thoroughly solid.

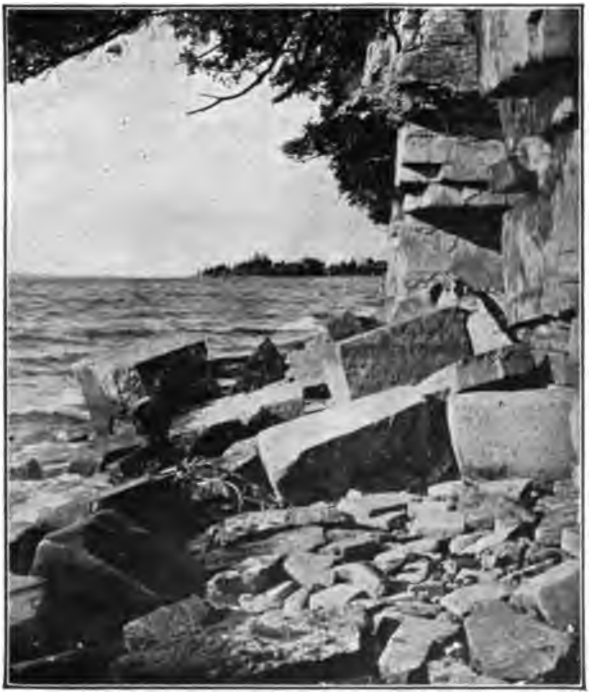

Fig. 301.—Angular blocks of rock which have fallen from the cliff above, as a result of undercutting by the waves. Grand Island, Lake Champlain. The rock is Black River limestone. Although from the shore of a lake instead of the sea, the principles illustrated are the same. (Perry.)

The impact of the waves is generally reinforced and made effective by the impact of the detritus they carry. The sand, the pebbles, and such stones as the waves can move are used as weapons of attack, being turned against one another and against the shore. Masses of rock too large for the waves to move (Fig. 301) are worn by the detritus driven back and forth over them, and in time reduced to movable dimensions (Fig. 302). They then become the tools of the waves, and in use, are reduced to smaller and smaller size. Thus bowlders are reduced to cobbles, cobbles to pebbles, pebbles to sand, and sand to silt. The silt is readily held in suspension in agitated water, and thus is carried out beyond the range of breakers, and settles in water so deep as not to be effectively agitated to its bottom. Thus one generation of bowlders after another is worn out, and the comminuted products are carried out from the immediate shore and deposited in deeper water.

The effectiveness of waves, whether they work by impact of water alone, or by impact of water and detritus, is dependent on their strength and on the concentration of their blows.[1] The strength of waves is dependent on the strength of the winds (or other generating cause) and the depth and expanse of the water, and the concentration of their blows is conditioned by the slope against which they break. On exposed ocean-coasts the fetch of the waves is always great. The winds are variable. For a given coast they have an average strength, but the effectiveness of wave-erosion is determined less by the average strength of waves than by the strength of the storm-waves. This is often very great. On the Atlantic and North Sea coasts of Britain, winter breakers which exert a pressure of three tons per square foot are not infrequent.[2] So great is the force of exceptional storm-waves that blocks of rock exceeding 100 tons in weight are known to have been moved by them. Ground-swells, "even when no wind is blowing, often cover the cliffs of north Scotland with sheets of water and foam up to heights of 100 or even nearly 200 feet. During northeasterly gales the windows of the Dunnet Head lighthouse, at a height of upwards of 300 feet above high-water mark, are said to be sometimes broken by stones swept up the cliffs by sheets of sea-water." [3] The average force of waves on the Atlantic coast of Britain has been found to be 611 lbs. per square foot in summer, and 2086 lbs. in winter.[4]

Where deep water extends up to the shore, the force of the wave is almost wholly expended near the water line; where shallow water borders the land, the force of the waves is expended over a greater area.

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[1] Willis. Jour. of Geol., Vol. I, p. 481.

[2] Stevenson. Trans. Roy. Soc. Edin., Vol. XVI, p. 25. Treatise on Harbors, p. 42. Quoted by Geikie, Text-book of Geology, p. 437.

[3] Geikie. Text-book of Geology, 3d ed., p. 437.

[4] Brit. Assoc. Rept., 1850, p. 26.

Waves are, therefore, most efficient on bold coasts bordered by broad expanses of deep water.

The less familiar phases of wave-work are accomplished by hydraulic pressure, compressed air, the use of ice, etc. When the water of a wave is driven into an open joint or a cave, the hydraulic pressure is great, and if the structure be weak, the rock may be broken. When water is driven with force into a cave, the compression of the air may be great if the wave be high enough to close the entrance. When the water runs out of a cave, the air within may be greatly rarefied, while that above exerts its normal pressure. In either case the roof of the cave, if it be weak, may be broken. At certain seasons of the year, especially during the spring, waves make destructive use of the ice which is then breaking up, but it is only in high latitudes that sea-ice is of consequence in this

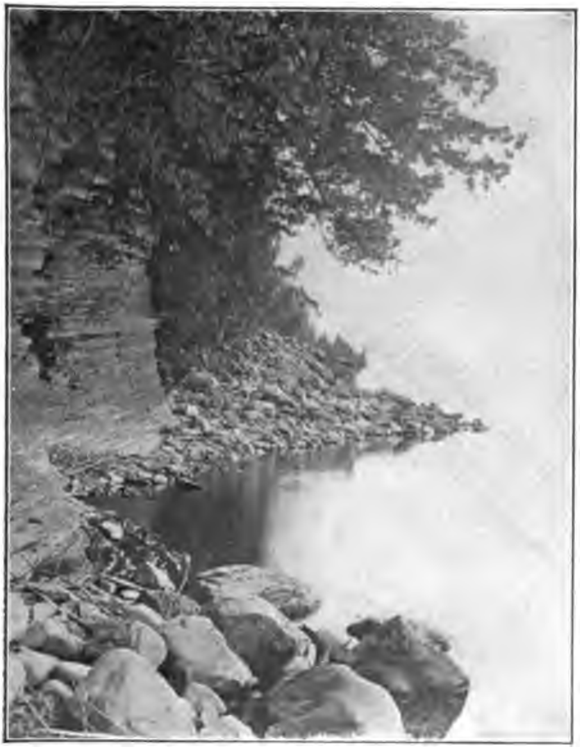


FIG. 302.—Showing blocks similar to those of Fig. 306, reduced and rounded by wave-action. Shore of Lake Champlain. The rock is Utica shale. (Perry.)

way. In general, the effect of its presence in keeping down waves overbalances its effect as an agent of erosion.

The direct effect of wave-erosion is restricted to a zone which is narrow both horizontally and vertically. There is no impact of breakers at levels lower than the troughs of the waves, though erosion may extend down to the limit of effective agitation (p. 341). The efficient impact of waves is limited upward by the level of the wave-crests, although the dash of the water produces feebler blows at higher levels.

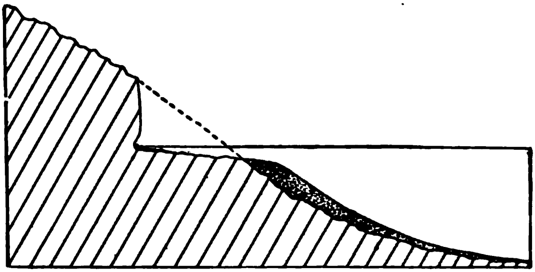


Fig. 303.—Diagram illustrating high sea-cliffs. It also shows a submerged terrace, due partly to wave-cutting (wave-cut terrace), and partly to building (wave-built terrace). (Gilbert.)

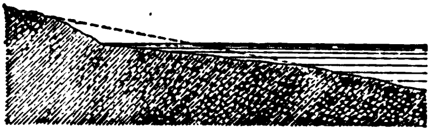


Fig. 304.—Diagram showing a low sea-cliff. (Gilbert.)

The rise and fall of the water during the flow and ebb of the tides gives the waves a greater vertical range than wind-waves alone would have. The vertical zone of direct wave-work is therefore limited above by the level of wave-crests, and below by the depth of wave-troughs (nearly). The indirect work of waves is limited only by the height of the shore, for as the zone of excavation is carried landward, masses higher up the slope are undermined and fall. The fallen rock temporarily protects the shore against the waves, but are themselves eventually broken up.

The pulsating current of the undertow (p. 341) has both an erosive and a transporting function. It carries the detritus of the shore to and fro, and dragging it over the bottom, continues downward the erosion initiated by the breakers. This downward erosion is the necessary concomitant of the shoreward progress of wave-erosion; for, if the land were merely planed away to the level of the wave-troughs, the incoming waves would break where shoal water was first reached, and become ineffective at the water margin. The rate of erosion by the undertow becomes less and less as the surface it affects is lowered. Littoral currents do little erosive work beyond that inflicted on the material which they transport.

The general result of wave-erosion is the advance of the sea on the land, the rate of advance being determined chiefly by the nature of the material attacked and the strength of the waves. Numerous as examples are of the retreat of coast-lines before the advance of the sea, it is not to be understood that the advance of the sea on the land is universal or uninterrupted. Numerous instances may be cited of the encroachment of the land on the sea. At Long Branch the advance of the sea, in spite of elaborate breakwaters, has been so rapid in recent years as to menace important buildings, while a few miles to the north and south,

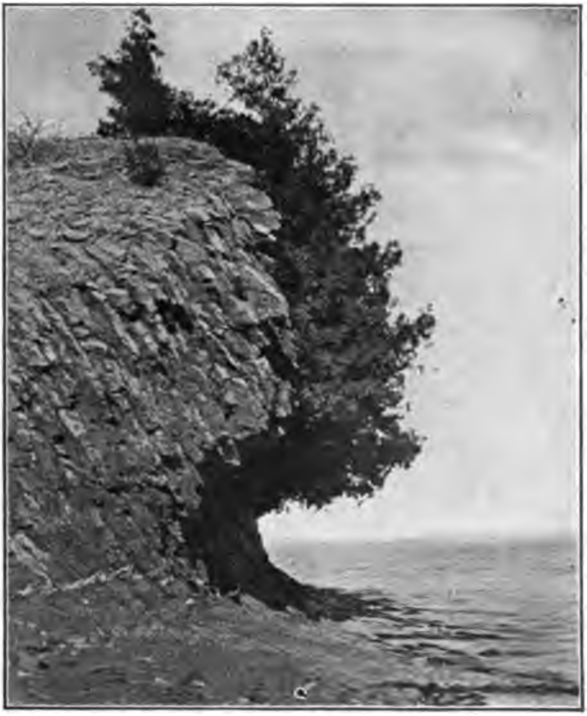


FIG. 305.—Steep cliff developed by waves. Allen Point, Grand Island, Lake Champlain. (Perry.)

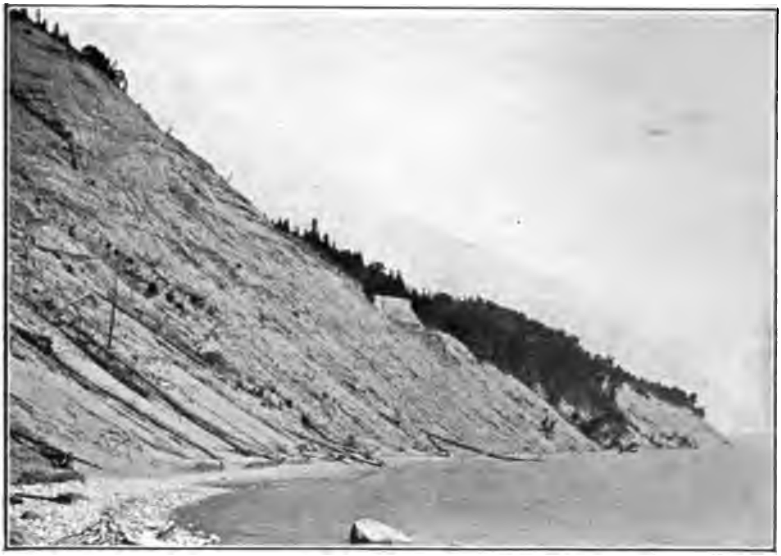


Fig. 306.—Cliff in unconsolidated material (bowlder clay), with lake-beach in foreground. South Manitou Island, Lake Michigan. (Russell, U. S. Geol. Surv.)

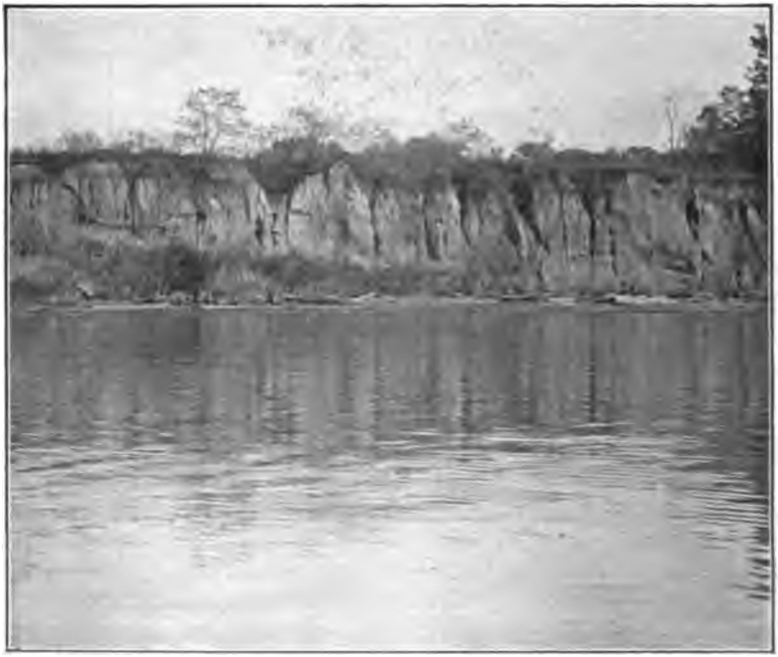
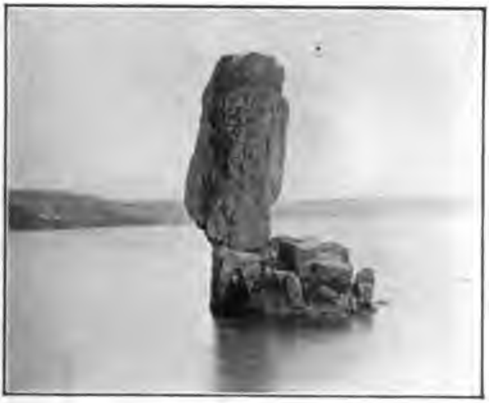


Fig. 307.—Steep cliff in unconsolidated material, the result of rapid cutting. Southeast extremity of Grove Point, Md.

the land is advancing in the face of the waves. The low coast of the Middle Netherlands has retreated two miles or more in historic times,[1] but the opposite tendency is shown at other points in the same region. On the coast of England the sites of villages have disappeared by the advance of the sea within historic times,[2] but the coast of the same island affords illustrations of land advance. On the south side of Nantucket island, the sea-cliff has been known to retreat before the waves as much as six feet in a single year.[3] Almost every considerable stretch of coast affords illustrations both of the advance of the sea on the land and of land on the sea; but in the long run, the former must exceed the latter, diastrophic movements aside.

Topographic Features Developed by Wave-erosion.

The sea-cliff.—The action of the waves, cutting as they do along a lefinite horizontal zone, has been compared to the action of a horizontal aw. As the waves cut into the shore at and near the water-level, the



. 308.—Standing Rock. A wave-erosion monument. West shore of Random Sound, south of Clarenville, N. F. (Walcott, U. S. Geol. Surv.)

terial above, being unsupported, falls, leaving a steep face above line of cutting. This steep face is known as *the sea-cliff* (Figs. 301 306). The same term is sometimes applied to the cliffs of lakes.

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[1] Davis. Physical Geography, p. 354.

[2] Dana. Manual of Geology, 4th ed., p. 219.

[3] Shaler. Sea and Land, p. 29.

The principles involved in the development of the sea-cliff are applicable to any broad stretch of water.

The height of the cliff depends on the height of the land on which the sea is advancing. Its slope may be steep or gentle (compare Figs. 303 to 306), according to the nature of the material of which it is composed and the rapidity of the cutting. Rapid cutting tends to produce steep cliffs and slow cutting gentle ones, for in the latter case weathering is more important relative to the cutting, and at sea-level (low altitudes) weathering generally tends to reduce the angle of slope. In general, the more resistant the material the steeper the slope of the cliff. Incoherent materials, such as sand and clay, are not likely to form steep cliffs; but if the cutting be very rapid, bold faces may be developed even in such materials (Fig. 307). If beds of slight resistance at sea-level underlie beds of greater resistance, the development of steep cliffs is favored. The structure of the cliff-rock also has an influence on the slope. The rock may be massive or bedded. If bedded, the beds may be horizontal, or they may dip at any angle, in any direction. The rock, whether stratified or not, may be abundantly or sparsely jointed. All these structures influence the slope and configuration of the sea-cliff (see Figs. 305 to 308).

Chimney-rocks, etc.—By working in along the joints of the rock, widening them and quarrying out the intervening blocks, pillars of rock ("chimney-rocks," "pulpit-rocks") or even considerable islets are sometimes isolated by the waves. This is most readily accomplished where the joints converge back from the shore. A well-known example of this sort is the "Old Man of Hoy" (Fig. 309) on the coast of the Orkneys. A pulpit-rock or other island, or any jutting point of rock may be pierced, giving an arch or bridge. La Roche Percée, a steep-faced isle near Gaspé Harbor, is an example.

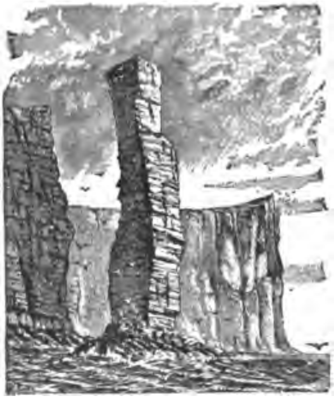

FIG. 309.—"Old Man of Hoy." (Geikie.)

Sea-caves. — Waves sometimes excavate caves at the bases of cliffs. This is especially likely to occur where the rock is much jointed and where the joints are not continued to the surface in a single plane. The bottom and roof of a sea-cave usually have a pronounced inclination landward. If the cliff be low, the cave may be extended landward until its roof is pierced. Through such an opening in the top of the cliff the water of the incoming waves may be forced in the form of spray. On the New England coast such holes are sometimes known as "spouting horns." Similar openings may be made, as already pointed out, by the compression or rarefaction of the air in the cave as the wave enters or retreats. If the roof of the cave be partially destroyed, the portion which remains may form an arch or *bridge*. Such a bridge occurs on Santa Cruz Island, California (Fig. 310).

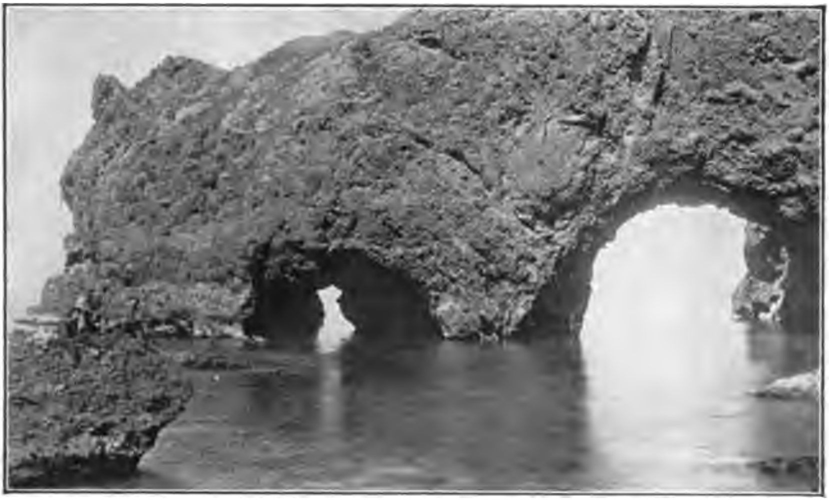

Fig. 310.—An arch developed by waves. Santa Cruz Island, Cal. (Law.)

The cave, the "spouting horn," the "bridge," the "pulpit-rock," and other isolated islets, are all closely associated with the sea-cliff in origin.

The wave-cut terrace.—The bottom of the sea-cliff is bordered by a submerged platform over which the water is shallow. This platform, or at any rate its landward portion, represents the area over which the water has advanced as the result of wave-cutting, and is, therefore, known as the *wave-cut terrace*. From the method of cliff development it will be seen that the wave-cut terrace is its necessary accompaniment. Such a terrace has a gentle slope to seaward, for its outer and older edge has been degraded longer and more. Its slope is influenced by the strength of the waves, being greater where they are stronger. The outer edge of the wave-cut terrace is often marked by an abrupt

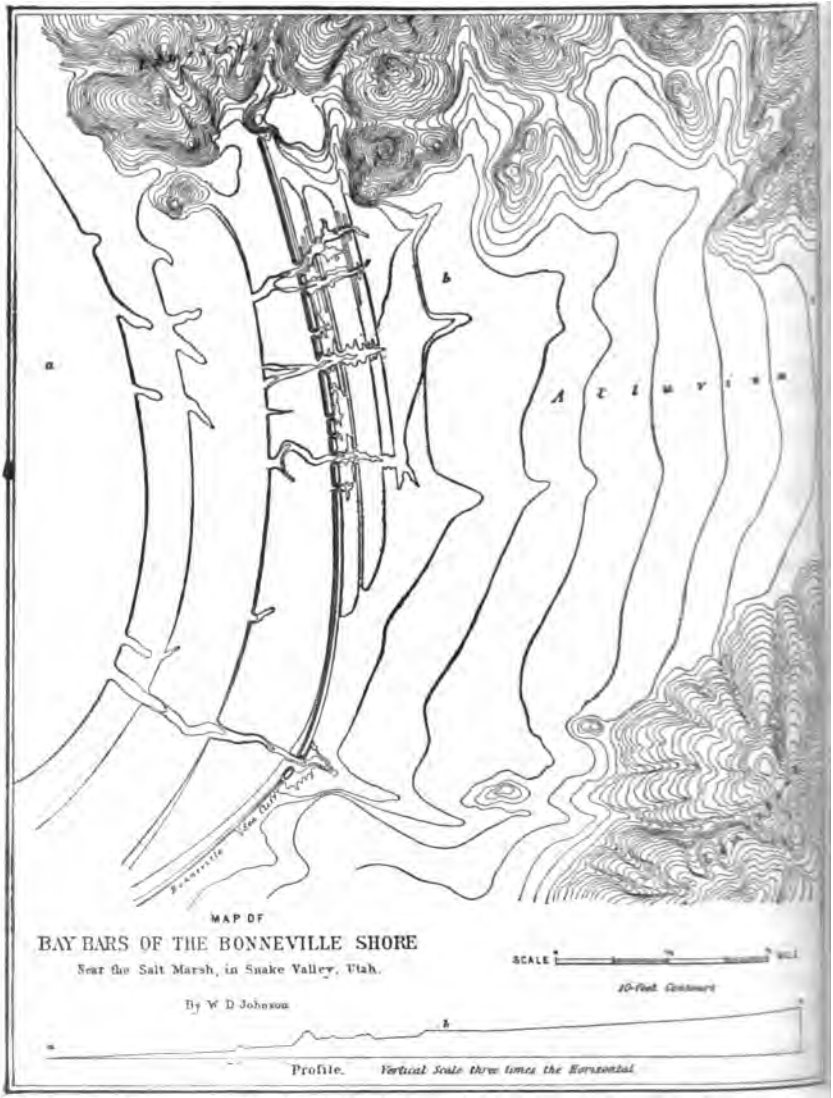
a
b
Alluvium
Bonneville Sea Cliff
MAP OF
BAY BARS OF THE BONNEVILLE SHORE
Near the Salt Marsh, in Snake Valley, Utah.
By W D Johnson
SCALE
10-Feet Contours
Profile. Vertical Scale three times the Horizontal

Fig. 311.—An elevated cliff above Great Salt Lake. In this case the water-level has been lowered. (Gilbert, U. S. Geol. Surv.)

descent. Fig. 303 represents the wave-cut terrace in its relation to the sea-cliff above.

So long as wave-cut terraces are submerged, they do not appear on topographic maps of the land, though they appear on the charts of the coasts; but if a coastal tract with wave-cut terraces be elevated, or if the sea-level be drawn down, the terraces become land. Elevated sea-cliffs and wave-cut terraces are among the best evidences of change of relative level between water and land (Fig. 311).

Wave-erosion and horizontal configuration.—The structure of the rock along shore has as much to do with the horizontal configuration of the wave-shaped coast, as with its relief. In general, waves develop reentrants in the less resistant portions of the shore, leaving the more resistant parts as headlands (San Pedro Point and Devil's Slide, Pl. XX, Coast of California). It is to be noted that the resistance of rock to wave-erosion is not determined by its hardness alone. Every division plane, whether due to bedding, to jointing, or to irregular fracture, is a source of weakness to the rock, and rock of great hardness may be so broken as to offer relatively little resistance. Inequalities of resistance, whatever their cause, give origin to inequalities of coastal configuration where wave-erosion is in progress. Given a coast of marked regularity and equal exposure, but composed of unequally resistant material, the waves will make it irregular by cutting most where the material is least resistant. A regular coast of uniform material, but unequal exposure, will be made irregular by the greater cutting at the points of greater exposure. A coast of marked irregularity and homogeneous material will be made more regular by the cutting off of the projecting points, because they are most exposed. With a given set of conditions, waves tend to develop a certain sort of shore-line which, so far as its horizontal form is concerned, is relatively stable. Such a shore-line may be said to be *mature*[1] so far as wave-erosion is concerned. Since coastal lands are, in general, both heterogeneous and unequally exposed, a mature coast-line is somewhat irregular. Its maturity is attained when the lesser exposure in the reentrants developed in the less resistant parts, balances the superior exposure of the projections of the more resistant portions.

Since the conditions of erosion along coasts are constantly, even if slowly, changing, maturity is constantly being approached, but rarely reached. Other forces and processes, such as those of aggradation, vulcanism, and diastrophism, are in operation along coasts, and their results are sometimes antagonistic to those of the waves. The horizontal configuration of coasts is, therefore, the result of many cooperating forces, of which waves are but one. It is, nevertheless, important to note the goal to which the waves are working, even though they are continually defeated in their attempt to reach it. Their immediate goal is an equilibrium of erosion-rate and maturity of configuration; their final goal is the destruction of the land and the deposition of its substance in the sea, that is, in a position nearer the center of gravity of the earth.

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[1] Gulliver, Shore Line Topography: Proc. Am. Acad. Arts and Sci., Vol. XXXIV, 1899, pp. 151–258. A valuable study of shore-line topography.

Transportation by Waves.

The material eroded from the shore by the waves in the shaping of the cliff and terrace is carried away by the joint action of the waves, undertow, and shore-currents.

The in-coming wave begins to shift material where it begins to drag bottom, that is, a little outside the line of breakers. From the line where transportation begins, to the line of breakers, bottom detritus is shifted shoreward by the waves, while the undertow tends to carry it back again. Between the breakers and the shore there is also a tendency for the on-shore movement to carry débris to the water's edge, and for the ebbing wave to carry it back again. The result of these opposed tendencies is to keep sediment in transit between the shore and the line of breakers. If the in-coming waves have a direction normal to the shore, the advance and recoil of the water move particles toward and from the shore, but effect no transfer along the shore; but the results which waves normal to the shore would achieve are always modified by other waves and by littoral currents.

If the in-coming wave is oblique to the shore, it shifts material in its own direction. The transfer by undertow, taken alone, would be sensibly normal to the shore, but the effect of the oblique waves is to slightly modify this direction. There is thus a slow transportation along shore, even in the absence of steady currents. A great amount of transportation would be effected in this way, though it would be carried on at a slow rate. Oblique waves also tend to develop a definite shore-current (p. 342) which affects both the amount and direction of the transportation. Any particle in suspension, or in motion on the bottom as the result of the wave or undertow, is shifted along shore by the littoral current, which affects the same water (Fig. 300). By the cooperation of wave- and shore-current, more and heavier material can be moved than by either alone, and the direction of movement is more nearly parallel to the shore than that of the wave. Similarly, by the cooperation of undertow and shore-current, more and heavier material can be moved than by either alone. The direction of movement is readily inferred from Fig. 300. The direction in which débris is shifted by wave- and shore-current is modified by the undertow, and the direction which would result from undertow and current is modified by the wave. It is often the waves of storms, rather than those of the prevailing winds, which determine the direction of greatest shore transportation.

The waves, the undertow, and the littoral currents work together in assorting the detritus of the shore. The coarsest parts may be beyond the power of all but the strongest waves. They accumulate where agitation is great. Less coarse parts are shifted farther from the site of greatest agitation, but no materials which are classed as coarse are carried beyond the depth of sensible movement. The coarse material which covers the bottom where the agitation of the water at the bottom is effective, constitutes *shore drift.*

Shore drift is not all derived from the shore by the cutting of the waves. A part of it is brought to the sea by streams and mingled with that eroded from the cliffs. The material which is fine enough to be held in suspension is measurably independent of depth. This is shown during storms when the water becomes turbid far beyond the line of breakers, and clears only after the waves have died away.

This sorting of shore drift, effected while it is in transportation, is often very perfect. The conditions favoring assortment are (1) vigorous wave-action, (2) prolonged transportation, and (3) a moderate volume of sediment.[1] The effect of these several conditions will be readily understood.

Extensive transportation of shore drift of a given degree of coarseness is favored by (1) strong waves and undertow, (2) continuous currents, and (3) shallow water, deepening but gradually off shore.

Deposition by Waves, Undertow, and Shore-currents.[2]

The beach.—The zone occupied by the shore drift in transit is the

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[1] Willis. Jour. of Geol., Vol. I, p. 481.

[2] See Gilbert. Topographic Features of Lake Shores, 5th Ann. Rept. U. S. Geol. Surv.

beach. The lower margin is beneath the water, a little beyond the line where the great storm-waves break. Its upper margin is at the level reached by storm-waves, and is usually a few feet above the level of still water. To the beach, material is brought from seaward by the in-coming waves, and from it detritus is carried out by the undertow. The cross-section of a beach is shown in Fig. 312. In horizontal position the beach follows the general boundary between water and land, though it does not conform to its minor irregularities (Fig. 313). The beach or barrier ridge often causes the deflection of the lower courses of streams descending to it (Pl. XXI).

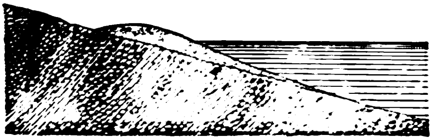


Fig. 312.—Cross-section of the beach. (Gilbert.)

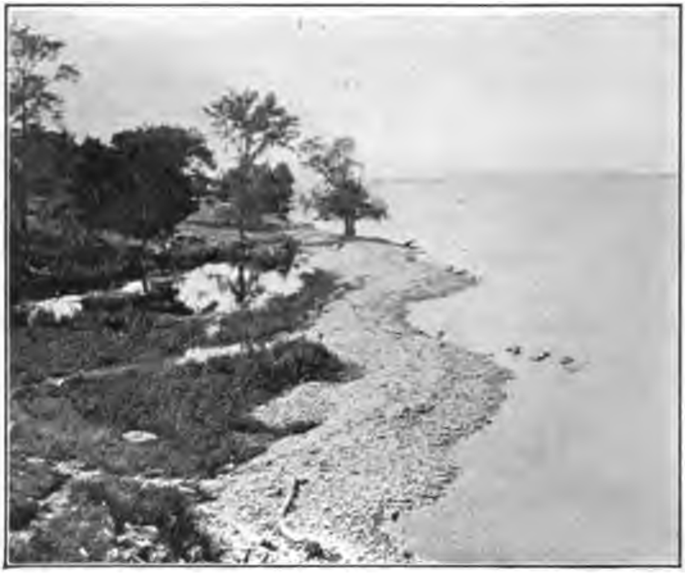
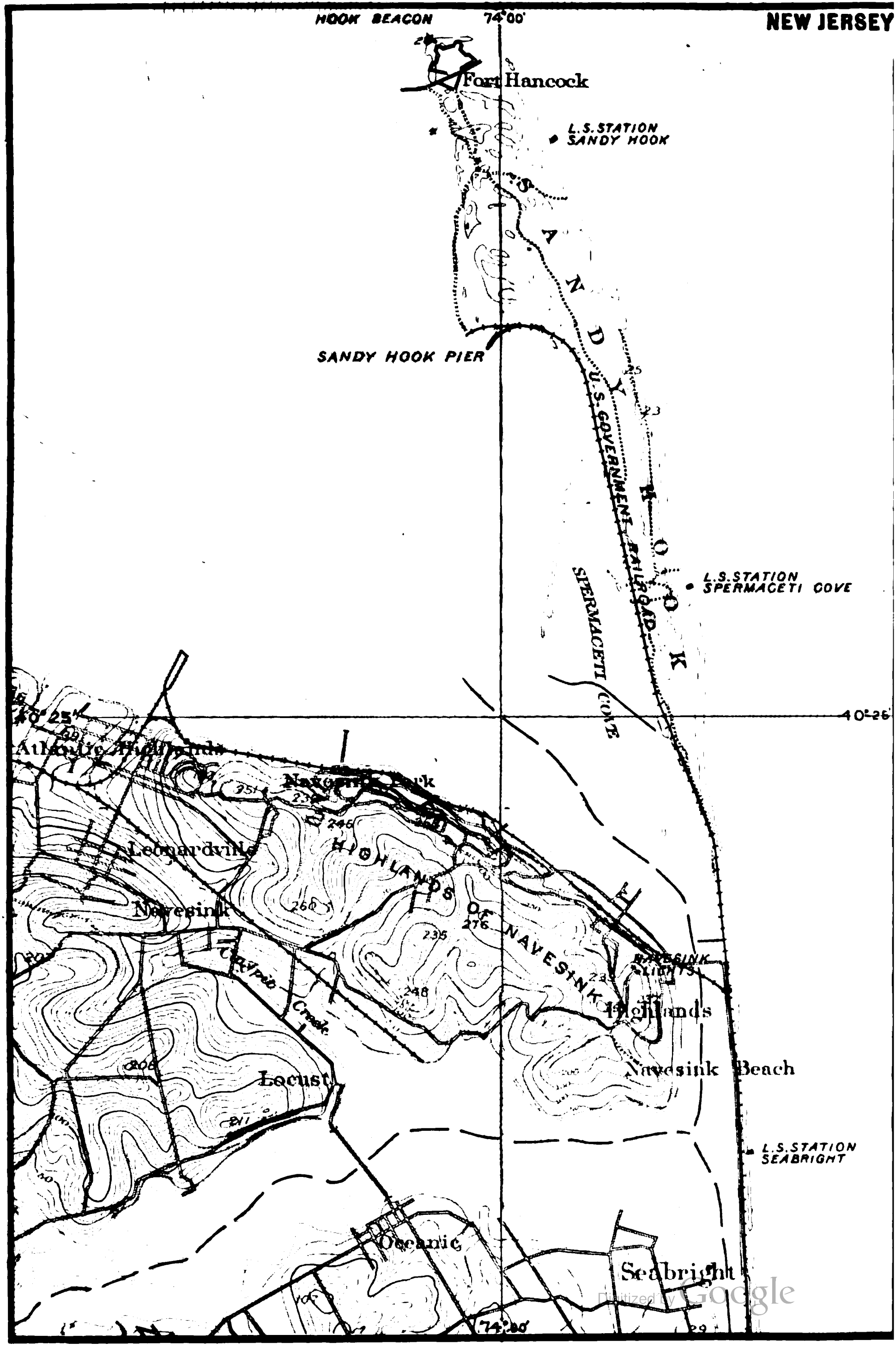


Fig. 313.—A lake-beach (barrier); Griffin's Bay, Lake Ontario.

The barrier.—When the agitation of the water along shore becomes insufficient to carry the material, it is dropped. In its deposition it assumes various forms. Where the bottom of the lake or sea near shore has a very gentle inclination, the in-coming waves break some distance from the shore-line, and it is here that the most violent agitation

PLATE XXI

NEW JERSEY



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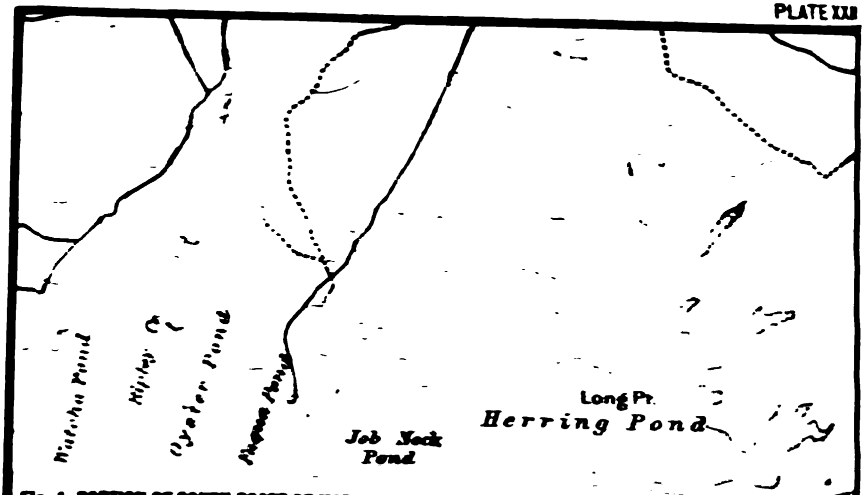

Fig. 1. PORTION OF SOUTH COAST OF MARTHAS VINEYARD, MASSACHUSETTS.

U. S. Geo Surv.

Scale 1+ mile per inch.

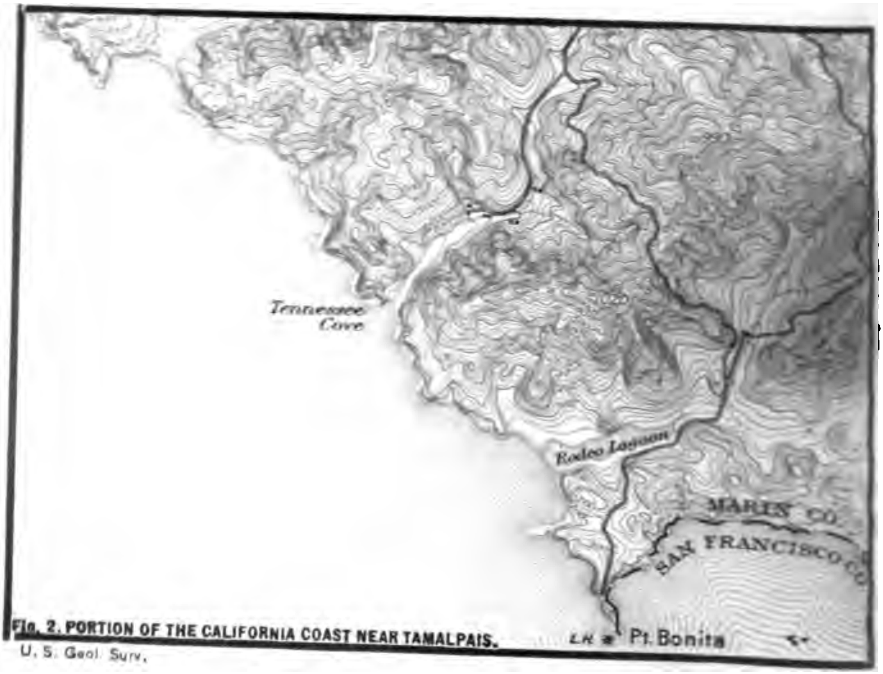

Fig. 2. PORTION OF THE CALIFORNIA COAST NEAR TAMALPAIS.

U. S. Geol. Surv.

Scale, 1+ mile per inch.

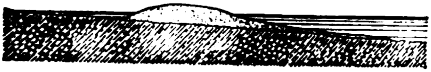

FIG. 314.—Section of a barrier. (Gilbert.)

occurs when the waves are strong. To this line of breakers, material is shifted from both directions: from shore by undertow, and from seaward by the waves. Accumulating here, it builds up a low ridge. This is a *barrier* (Fig. 314). If it is built up above the surface of the water by storm-waves, it may shut in a lagoon behind it, and this may ultimately be filled by sediment washed down from the land. At one stage in the filling, the lagoon becomes a marsh.[1] In the part which the barrier plays in the history of a coast, it is identical with the beach.

The spit, the bar, and the loop.—The disposition of shore-deposits depends largely on the currents at and near shore. If the coast-line is deeply indented, the littoral current usually fails to follow the reentrants. In holding its course across the mouth of a small bay, a shore-current usually passes into deeper water. Here its velocity is checked because its motion is communicated to the water beneath it, and a larger amount of water being involved in the motion, the motion of each part is diminished. If sediment was being moved along its bottom before the current was checked, some part of it is dropped when and where the current is slackened. It follows that deposition commonly takes place beneath a littoral current as it crosses the mouth of a bay. The belt of deposition is often narrow, and the result is the construction of a ridge beneath the water in the direction of the current. The current would never build the embankment up to the water-level, but when

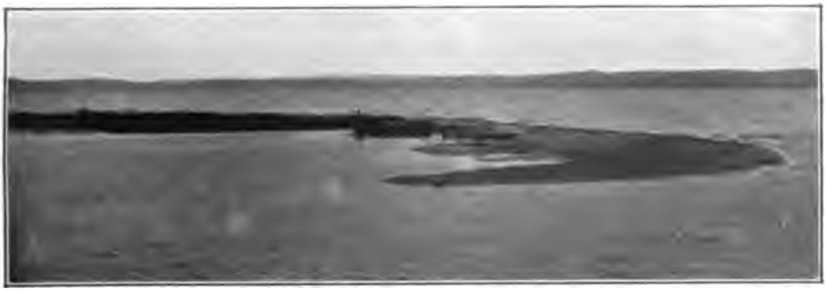

FIG. 315.—A recurved spit. Dutch Point, Grand Traverse Bay, Lake Michigan.

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[1] Shaler, Sea Coast Swamps of the U. S., 6th Ann. Rept. U. S. Geol. Surv.; and Merrill, Pop. Sci. Mo., Oct., 1890.

its surface approaches the level of effective agitation, the waves may begin to work on it, as on a barrier, and may build it up to, and even above, the surface of the water. So long as the end of such an embankment is free, it is a *spit* (Fig. 315 and Pl. XXI). If the spit be lengthened until it crosses, or nearly crosses, the bay, shutting it off from the open water, it becomes a bar. Bars have shut in lakes (ponds) on the coast of Martha's Vineyard, Mass. (Fig. 1, Pl. XXII), and lakes and lagoons at numerous points both on the Atlantic and the Pacific coasts (Fig. 2, Pl. XXII, Rodeo lagoon). The same phenomena are to be seen along many lake shores. Bars sometimes tie islands to the mainland (Pl. XXIII, Fig. 1, Nahant, Mass.; Fig. 2, near Biddeford, Me.). The structure of a bar as seen in cross-section is shown in Fig. 316.

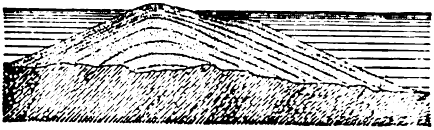


FIG. 316.—Cross-section of a bar. (Gilbert.)

The construction of a spit has been aptly compared to the construction of a railway embankment across a depression. The material is first carried out from the bordering upland (shallow water) and dumped where the slope to the depression (deep water) begins. The embankment thus begun is extended by the carrying out of new material, which is left at the end of the dump already made.

If the bay across which the bar is built receives abundant drainage from the land, the outflow from the bay may be sufficient to prevent the completion of the bar (Fig. 2, Pl. XXII), for when the growth of the spit has sufficiently narrowed the outlet of the bay, the sediment brought to the end of the spit by the littoral current will be swept out beyond the spit by the current setting out from the bay.

The completion of a bar may be interfered with by tidal currents, even without land-drainage. Currents generated by the tides may sweep in or out of the bay with increased force as the entrance is narrowed, carrying in or out the sediment which the littoral current would have left at the end of the spit. The scour of the tides often insures deep entrances (inlets) to bays, and maintains definite channels or "thorofares" in the lagoon marshes behind barriers and spits. The sediment brought down from the land, as well as that washed in by tidal currents and waves, tends to fill up the lagoon behind a barrier, a spit, or a bar, converting it into land (Fig. 317).

Since spits and bars are built only where there is shore-drift in transit, they are always built out from a beach or barrier. The distal end of the

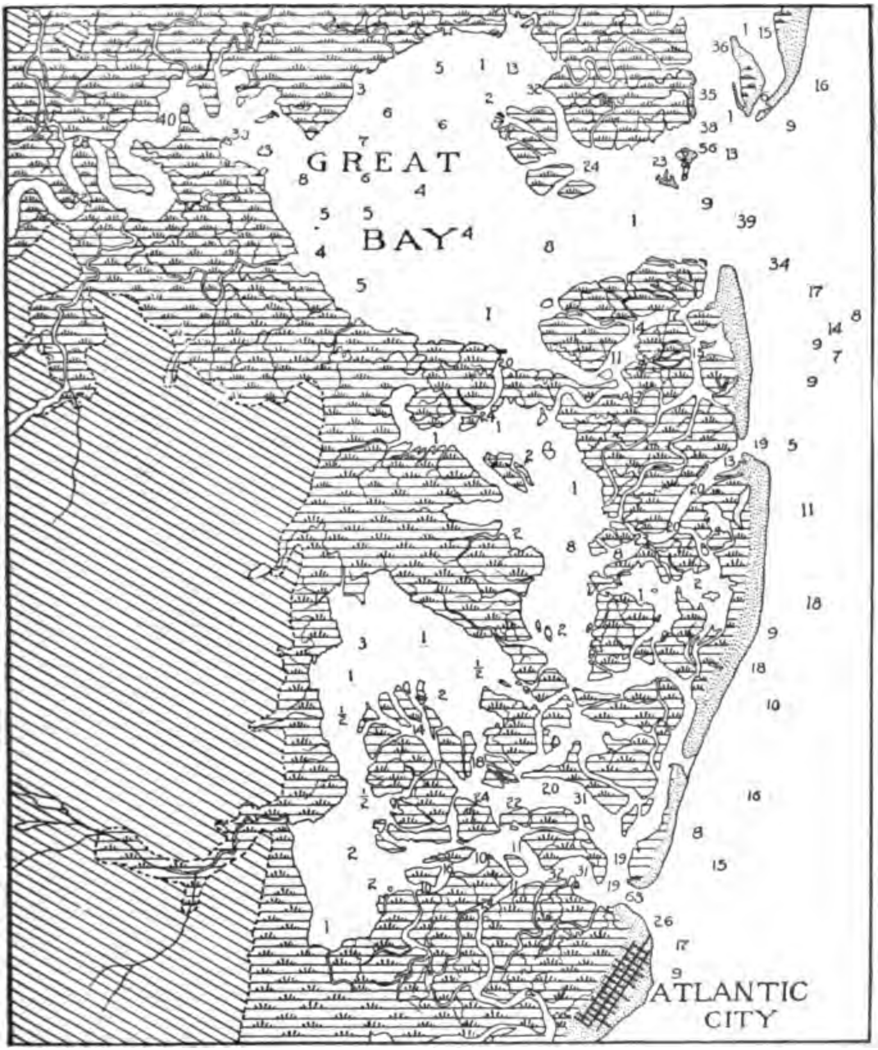


Fig. 317.—Sketch of a portion of the New Jersey coast. The dotted belt next the sea is the barrier, modified by the wind. The area marked by the diagonal lines is the mainland. In the marshy area between, there are numerous channels or "thorofares" kept open by the currents. The figures show the depths of water in feet. Scale about ⅜ inch = 1 mile.

bar may also join a beach or barrier. Traced back to its source, the beach from which a spit leads out is often found to terminate in the cliff from which the material of the beach and the spit were derived (Pl. XX and Fig. 2, Pl. XXII). In such cases the sediment of the beach has

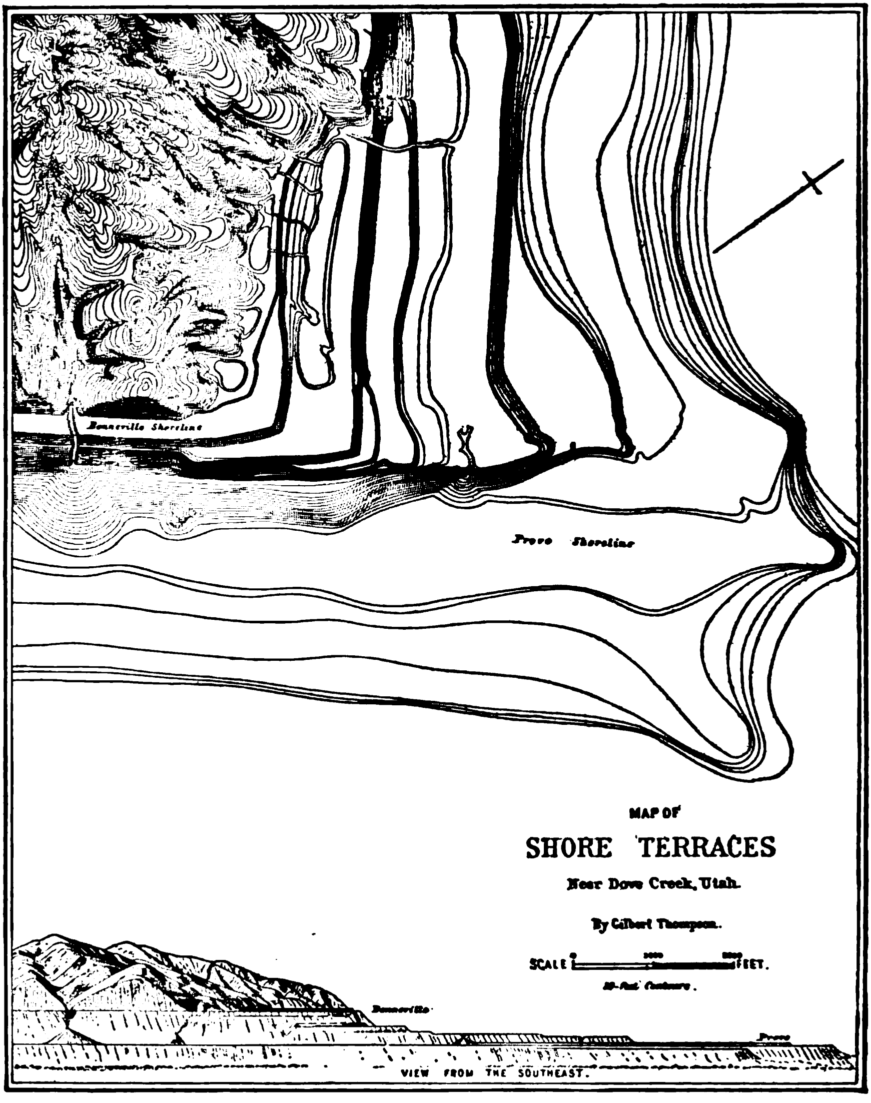


FIG. 318.—Map of shore-terraces, largely wave-built. Lake Bonneville. (Gilbert.)

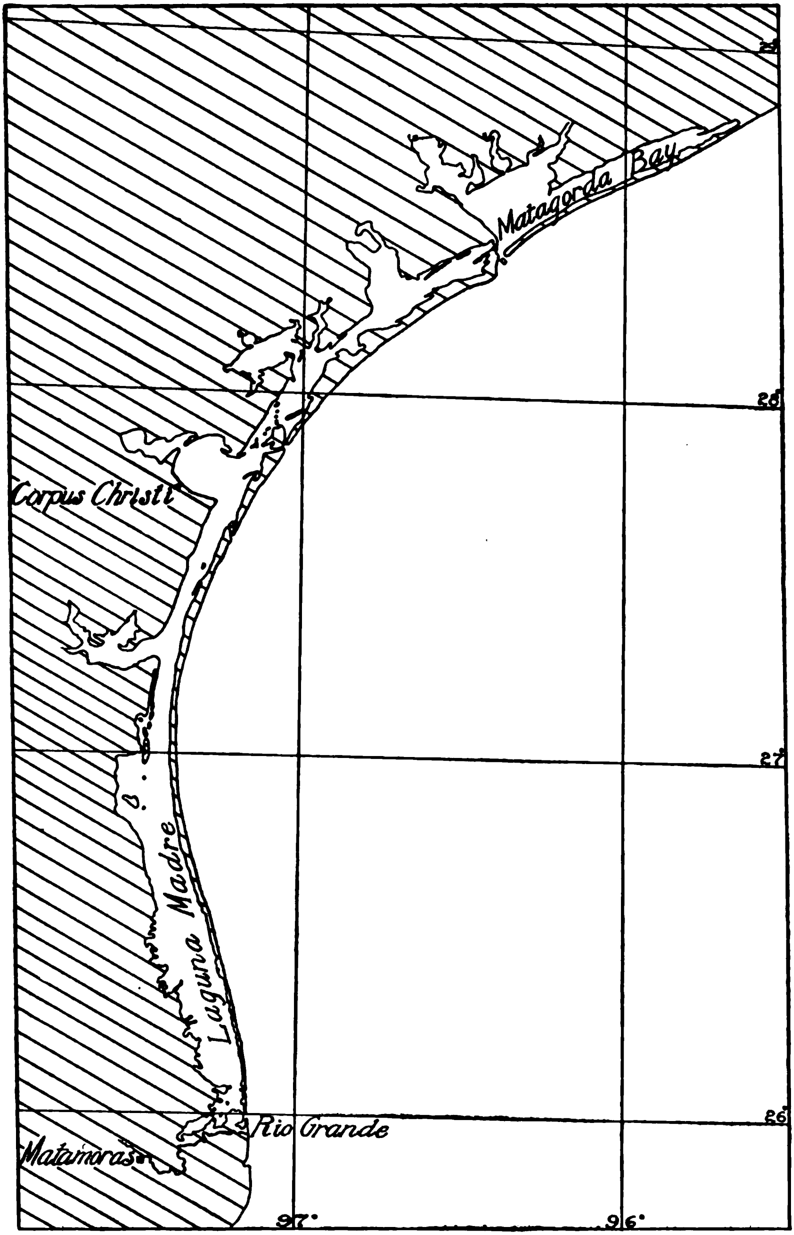


FIG. 319.—A portion of the Texas coast showing the tendency of shore-deposition to simplify the coast-line. The deposits (the narrow necks of land parallel to the coast) shut in the bays. (From chart of C. and G. Surv.)

been shifted but a short distance; but in other cases it has traveled far.

The spit is usually either straight or in conformity with the general course of the shore-current, but since the littoral current itself is subject to alteration as the result of shifting winds, the spit may depart

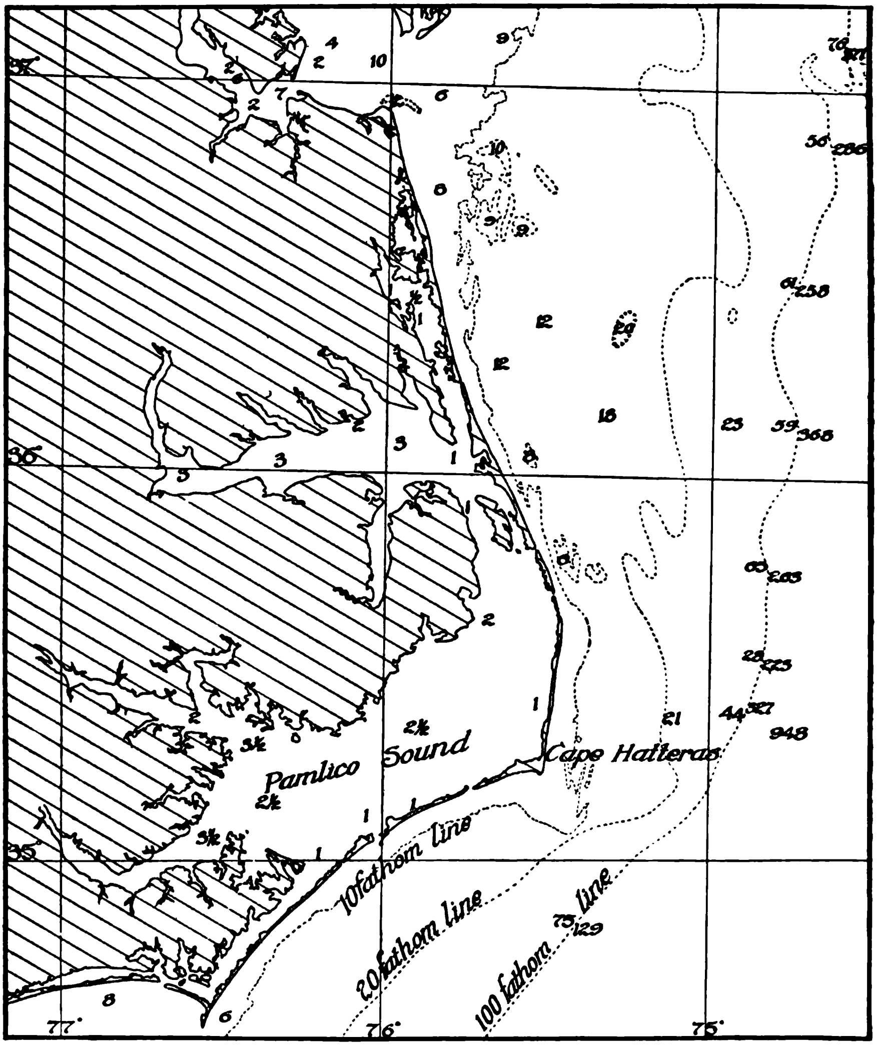


FIG. 320.—Map showing that in the early stages of the simplification of a shore-line the irregularities are increased. The numbers indicate the depth of water in fathoms. (From chart of C. and G. Surv.)

from straightness. Winds which simply reverse the direction of the littoral current retard its construction, but may not otherwise affect it; but if a strong current be made to flow past the end of a spit, it may cut away its extremity and rebuild the materials into a smaller spit, joining the main one at an angle. This gives rise to a *hook* (Fig. 315). Successive storms may develop successive hooks along the side of a growing spit. The end of a hook may be so extended as to join the mainland, when it becomes a *loop*.

Wave-built terraces.—Under the influence of off-shore currents, littoral currents may be drawn from the coast-line. If such a current continues as a well-defined surface-current, it builds a spit, but if it spreads, it tends to build a terrace. The accumulation then is not at the end of a beach, as in the case of a spit, but on its side, and the result of the deposition is to carry the beach seaward. The undertow abets the process. The widened beach is a *wave-built terrace*. The wave-built terrace often borders the wave-cut terrace along its seaward margin (Figs. 303 and 318). With the help of waves, the surface of the terrace may be built up into land by the expansion of the crest of the beach. Terrace-cutting and terrace-building are both involved in the development of the continental shelves.

Beach ridges, spits, bars, etc., like sea-cliffs and wave-cut terraces, are often preserved after the relative level of sea and land has changed. If the shore has risen, relatively or absolutely, these features are relied on as evidences of the change. If shore features be submerged instead of elevated, they furnish less accessible, though not less real, evidence of the change of level. Similar features about lakes have a like significance, but in this case it is often demonstrable that it is the water rather than the land which has changed its level.

Effect of Shore-deposition on Coastal Configuration.

The tendency of shore-deposition is to cut off bays and to straighten and simplify the shore-lines. This is abundantly illustrated along the Atlantic and Gulf coasts of the United States (see Fig. 319 and Pl. XXII). It is to be noted, however, that in the simplification of the shore-line through deposition, the initial stages often result in great irregularity (Fig. 320 and Pl. XXIII). In some cases, the irregularities are not temporary. Thus deltas (p. 198), though not wholly the work of sea- (or lake-) water, often constitute irregularities of a more or less permanent nature. This is the case where they project beyond the general trend of the coast-line. Where, on the other hand, they are built at the heads of bays, they tend to simplify the coast-line by obliterating the indentation. The delta at the head of the Gulf of California is an example. So too is the delta of the Mississippi, the real head of which is far above the present debouchure of the stream. The form of the delta in ground-plan depends on the horizontal configuration of the coast where it is developed, on the strength of the waves and shore-currents, and on their relation to the amount of detritus contributed by the stream concerned. Good illustrations are furnished by the Gulf of Mexico where the deltas of the Mississippi and Rio Grande are in contrast.

So far as concerns the vertical configuration of coasts, erosion and deposition are in contrast, for while the former tends to develop steep, irregular, and often high slopes (p. 349) from the land to the sea, the latter tends to develop gentle, regular, and low ones. A partial exception to the latter part of this general statement comes about through the building of dunes, the material for which is furnished by the waves.

SUMMARY OF COASTAL IRREGULARITIES.

The horizontal irregularities of coasts are both large and small. Some of them, like Florida, Sandy Hook, etc., consist primarily of projections of land into the sea; others, like Chesapeake Bay, the Gulf of Mexico, and Puget Sound, are projections of the sea into the land; while still others, like the Gulf of California and its associated peninsula, cannot readily be put in either of the foregoing classes. Some of the irregularities of the land border, such as Yucatan, are more or less nearly normal to the general trend of the coast which they affect, while others, such as the "beaches" along the Atlantic and Gulf coasts of the United States (Figs. 319 and 320), are more or less nearly parallel with it. Some of the irregularities, especially some of the small ones, are more or less angular in their outline (Pl. XX and parts of Fig. 2, Pl. XXII), while others are bounded by curves instead.

In many cases more than one factor has been involved in the development of irregularities. In the case of great irregularities, diastrophism

PLATE XXIII.

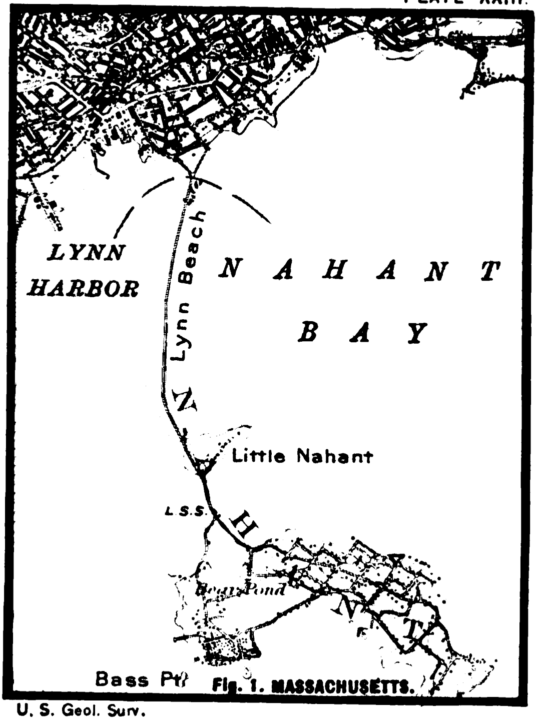


Fig. 1. MASSACHUSETTS.

U. S. Geol. Surv.

Scale, 1+ mile per inch.

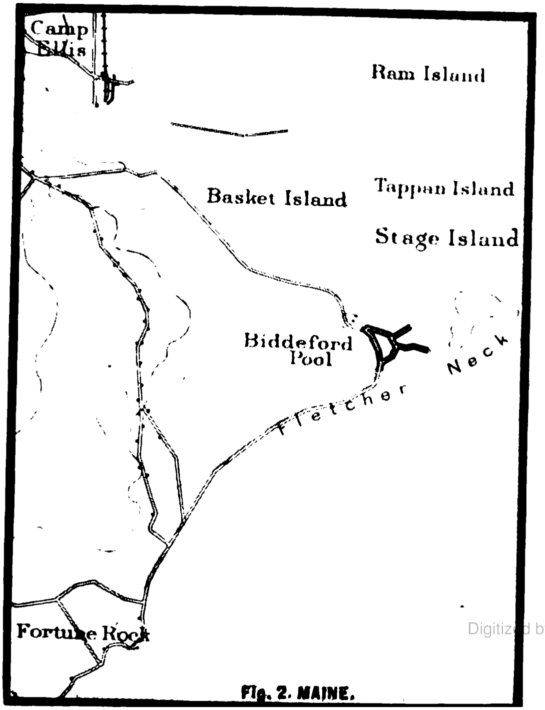
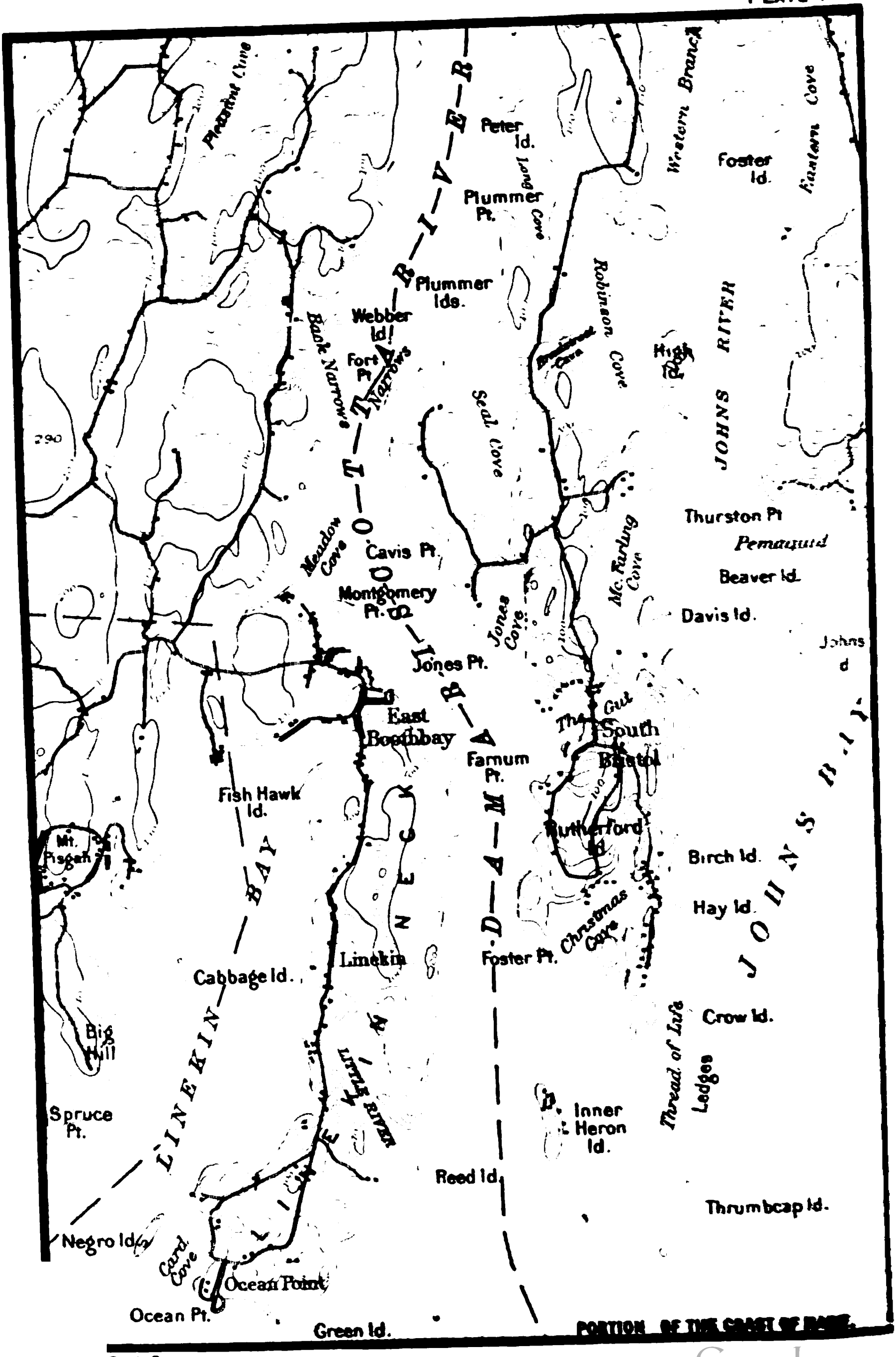


Fig. 2. MAINE.



PORTION OF THE COAST OF MAINE.

Geol. Surv.

Scale, 1+ mile per inch.

has generally been the dominant factor. The Gulf of Mexico and the Mediterranean Sea perhaps represent differential subsidence, while Florida and the Iberian peninsula represent differential uplift (relative, though perhaps not absolute). The narrow bays which indent many coasts generally represent the subsidence of a region previously affected by valleys (Fig. 297). Many of them, such as Narragansett, Delaware, and Chesapeake Bays, are primarily the drowned ends of river valleys, while others, such as Puget Sound,[1] are primarily structural valleys (synclines). Many of the long and narrow bays or fiords common in the high latitudes of North America and Europe (Fig. 266, p. 293) appear to be the drowned ends of valleys previously deepened by glaciers. The drowned ends of river canyons, and the submerged parts of valleys excavated (not sunk) beneath the sea by glaciers, would also be fiords.

The processes which develop coastal indentations, together with the antecedent subaërial and the subsequent wave gradation, account for most of the islands which affect indented coasts. Some of them are high and some low for reasons which will be readily understood. The long narrow belts of land constituting irregularities parallel to the general trend of the coast (Figs. 319 and 320) are usually the result of deposition in shallow water. They are usually sand or coral reefs, built up above water-level by waves. The deposits at the debouchures of streams give rise to projecting deltas. Most small irregularities of angular form, especially if high (Pl. XX), indicate wave-erosion, and their details of form are determined by the structure of the rock along shore, while most irregularities of curved outline involve something of shore-deposition, if not due wholly to it. Glaciation, or glaciation and subsidence, may also give rise to peninsulas, capes, and islands of curved outlines (Pl. XXIV, coast of Maine). Curving outlines may, however, be developed by erosion alone in weak rock structures. This is illustrated by the weak rock structures (clay, sand, etc.) of most of the Atlantic coastal plain. Thus inspection of the horizontal configuration of coasts will often indicate the processes which have been dominant there in recent times. On the other hand, the interpretations of many coastal irregularities, such as Hudson Bay, Puget Sound, the Gulf of California, the Baltic Sea, etc., are not to be read from the map. In such cases, diastrophism and gradation have usually cooperated, but the relative importance of the two processes can only be determined by detailed study in the field. When it is remembered that the tendency of shore-erosion is to reduce great irregularities of horizontal configuration, though not to obliterate small ones if the coast be heterogeneous in composition (p. 353), and that the tendency of shore-deposition is also to regularity, it is clear that the great irregularities of coast-lines are due neither to shore-erosion nor to shore-deposition, though minor ones may be due to either.

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[1] Willis. Bull. Geol. Soc. Amer., Vol. IX, p. 113, and Tacoma, Wash., Folio, U. S. Geol. Surv.

THE WORK OF OCEAN-CURRENTS.

As agents of erosion, ocean-currents are not, in general, of great importance. Currents which reach the bottom are comparable, in their effects, to rivers of the same velocity and volume; but most ocean-currents do not touch bottom, and, therefore, do not erode it. Where the current agitates the bottom sensibly, as it often does in shallow water, the bottom is abraded, and in the lee of such places it is doubtless aggraded. Since ocean-currents do not, for the most part, flow in shallow water, their erosive work is, on the whole, relatively slight; but where they are forced through narrow and shallow passageways, their abrasive work may be considerable. Thus the Gulf Stream, where it issues from the Gulf, has a velocity of four or five miles per hour, and its shallow and narrow channel is current-swept.

A rough test of the abrasive work of an ocean-current is found in the nature of the bottom beneath it. If this be hard, it indicates that the loose sediment on the floor of the ocean has been swept away, while the presence of fine detritus indicates that the current is not wearing. Thus the abrasive power of the Gulf Stream is known to continue somewhat beyond its narrow channel, for on the Blake plateau (between the Bahamas and Cape Hatteras), where the water is 600 fathoms and less in depth, "the bottom of the Gulf Stream . . . is swept clean of lime and ooze and is nearly barren of animal life."[1] Other illustrations of the erosive power of currents have been noted near Gibraltar in water 500 fathoms deep, and between the Canary Islands at depths of 1000 fathoms.[2] In spite of these examples, and of many others which probably exist in similar situations, it yet remains true that ocean-currents are on the whole but feeble agents of erosion.

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[1] Agassiz. Three Cruises of the Blake, Vol. I, p. 259. Agassiz would ascribe the Blake plateau itself to the Gulf Stream, p. 138. See also Am. Jour. Sci., Vol. XXXV, 1888, p. 498.

[2] Reade. Phil. Mag., Vol. XXV (1888), p. 342.

As agents of transportation, ocean-currents are scarcely more important than as agents of corrasion, for they transport only what they erode, if the life which inhabits them be left out of consideration. This phase of their work has probably been exaggerated through a confusion of transporting energy and actual transportation. Ocean-currents which do not touch bottom roll no sediment and carry only what may be held in suspension. A river's power of transporting sediment in suspension is due largely to the cross-currents occasioned by the unevenness of its resistant bottom (p. 117). If a particle of mud in suspension in a river drops to the bottom, as it frequently does, it may be picked up again and carried forward. If, on the other hand, a particle in suspension in an ocean-current once escapes the moving water by settling through it, the current which does not drag bottom has no chance to pick it up again. Very fine sediment may be carried by an ocean-current far beyond the point where it was acquired, but currents which do not touch bottom are rarely strong enough to hold any but the finest material for any considerable length of time. As transporters of sediment, therefore, ocean-currents are at a great disadvantage as compared with rivers.

How readily particles of extreme fineness may be kept in suspension, and how little agitation is necessary to keep them from sinking, is shown by the experiments of Sorby, who showed that while a sand grain $\frac{1}{100}$ of an inch in diameter will settle one foot per second in still water, fine particles of clay require days to sink through the same distance. The Challenger found fine sediment derived from the land 400 miles from the coast of Africa, and that not opposite the debouchure of any large river. Sediment settles more readily in salt water than in fresh, despite the fact that the former is heavier. This is presumably because the salt diminishes the cohesion of the water.

Deposition by ocean-currents is limited by their transportation. Only where they erode their bottoms do they gather coarse materials, and only in the lee of such places are their deposits coarse. Since the material which they carry is generally fine, it is widely distributed before deposition.

Ocean-currents have little influence on the configuration of coast-lines.

DEPOSITS ON THE OCEAN-BED.

Something has already been said concerning the sediments which accumulate in the shallow waters along shores; but the area of marine sedimentation is as extensive as the ocean itself, and the deposits must now be reviewed from another point of view.

Oceanic deposits may be conveniently divided into two chief groups, dependent on the depth of the water in which they are made.[1] These groups are (1) *shallow-water deposits*, made in water less than some such depth as 100 fathoms, and (2) *deep-sea deposits*, laid down in water of greater depth. The selection of the 100-fathom line as the dividing depth is less arbitrary than it seems, for passing outward from the shore, it is at about this depth that the bottom ceases to be commonly disturbed by the action of currents and waves; that sunlight and vegetable life cease to be important at the bottom; and that the coarser sediments which predominate along shore give place, as a rule, to muds and oozes. Furthermore, the 100-fathom line (or some line very near it) is an important one in the physical relief of the globe, for it appears to mark, approximately, the junction of continental plateaus and ocean-basins. Only because the latter are a little over-full does the water run over their rims, covering about 10,000,000 square miles of the borders of the continents, converting them from land into epicontinental seas.

Aside from the deposits made by organisms, shallow-water deposits are divisible into two groups—(*a*) those immediately along the shore, the *littoral deposits*, and (*b*) those made between the littoral zone and the 100-fathom line. Both are terrigenous. The deep-sea deposits likewise are divisible into two groups, (*a*) *terrigenous deposits* formed close to land, and made up chiefly of materials derived immediately from the disintegration of land formations; and (*b*) the *pelagic deposits*, made up chiefly of the remains of pelagic organisms and the ultimate products arising from the decomposition of rocks and minerals. The former predominate in the less deep waters relatively near shore; the latter in the deeper water far from land. The shallow- and deep-water deposits grade into each other in a belt along the 100-fathom line.

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[1] Murray. Challenger Report, Deep Sea Deposits, pp. 184, 185.

Shallow-water Deposits.

Littoral deposits.—The littoral zone is the zone between high- and low-water marks. It is the zone in which bowlders, gravels, sands, and all coarser materials accumulate, though muds are occasionally met with in sheltered estuaries. Generally speaking, the nature of these deposits is determined by the character of the adjoining lands and the nature of the local organisms. "The heavier materials brought by rivers from high terrestrial regions, or thrown up by the tides and waves of the sea, are here arranged with great diversity of stratification through the alternate play of the winds and waves. Twice in the twenty-four hours the littoral zone is covered by water and exposed to the direct rays of the sun or the cooling effects of the night. There is a great range of temperature; mechanical agencies produce their maximum effects,"[1] and physical conditions in general are most varied. Still greater diversity is introduced by the fact that the zone is inhabited by both marine and terrestrial organisms, while the evaporation of the sea-water which flows over tidal marshes and lagoons leads to the formation of saline deposits. If the length of the coast-lines of the world be taken at 125,000 miles (about 200,000 kilometers), and the average width of this zone at half a mile, these deposits are now forming over an area of 62,500 square miles (about 160,000 square kilometers) of the earth's surface.

Non-littoral, mechanical deposits in shallow water.—These deposits are laid down in the zone of the ocean between low-water mark and the 100-fathom line. They cover about 10,000,000 square miles.[2] Their composition is much the same as that of the littoral deposits, with which they are continuous, though on the whole they are finer. At their lower limit they pass insensibly into the fine deposits of the deep sea. Coarse material, such as gravel and sand, prevails, though in special situations, such as depressions and inclosed basins, muddy deposits are found. While some of the deposits are wholly composed of inorganic débris, organic remains are freely mingled with others. The mechanical effects of tides, currents, and waves are everywhere present, but become less and less well marked as the 100-fathom line is approached. The forms of vegetable and animal life are numerous, though the former decrease as depths which exclude the sunlight are approached.

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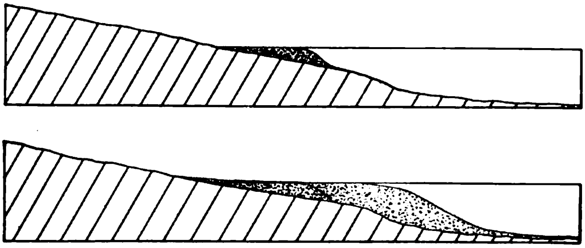
[1] Murray, loc. cit., pp. 187, 188.

[2] Ibid.

Both littoral deposits and deposits in shallow water outside the littoral zone have already been referred to in connection with the work of waves and currents (pp. 355–66). A few additional points only need here be added.

In general the coarser sediments are lodged near shore and those farther from the land become progressively finer. Even the coarser part of the material carried in suspension by the undertow is partly left in the shallow water. On the other hand, waves of exceptional strength may carry coarse material into water of some depth. Thus coarse shingle (gravel) and even bowlders have been found at depths of 10 fathoms.[1] Coarse deposits may extend far out from land if the waves are strong, and especially if the water is shallow, and since the zone of shallow water may be extended seaward by the aggradation of the bottom, shallow-water deposits may cover extensive areas. They may become deep at the same time, for as the outer border of the shallow-water zone is shifted seaward by aggradation, the vertical space to be filled becomes greater (compare Figs. 321 and 322).



FIGS. 321 and 322.—Diagrams showing how shallow-water deposits may attain considerable depth by the shifting of the zone of deposition seaward.

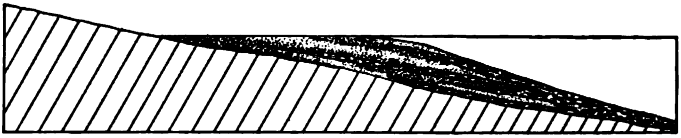


FIG. 323.—Diagram showing the interwedging of gravel-, sand-, and mud-beds.

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[1] Stevenson. Harbors, 2d ed., p. 15.

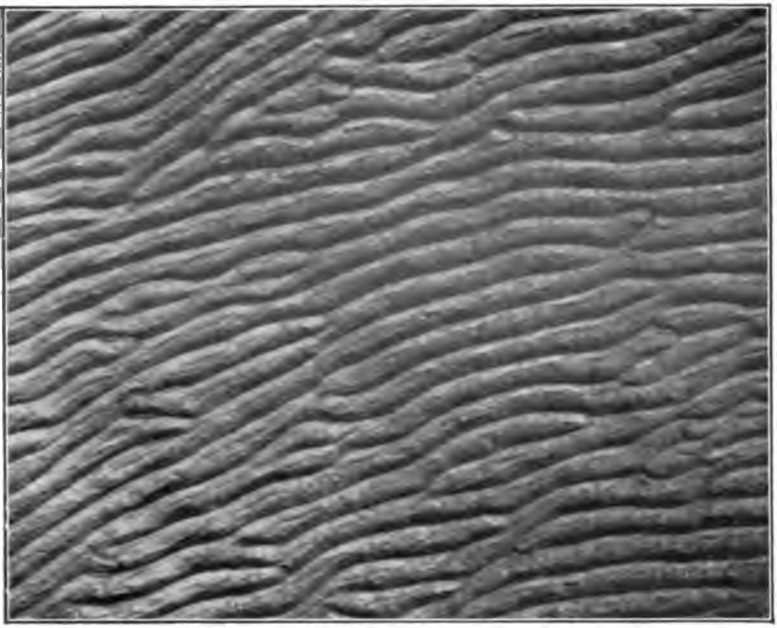

Fig. 324.—Ripple-marks.

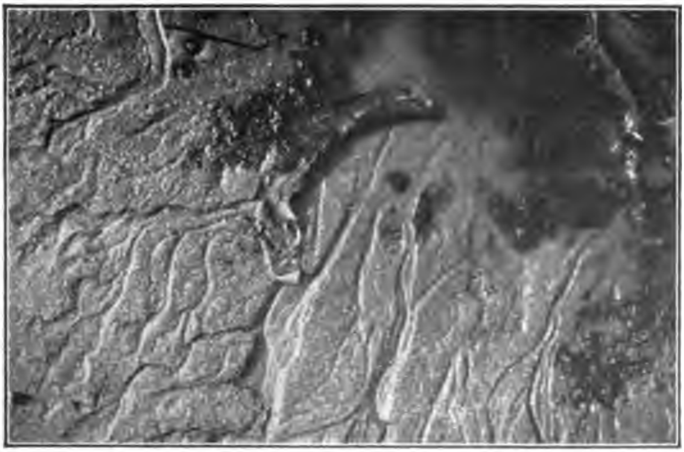

Fig. 325.—Rill-marks resembling impressions of seaweeds. Beach at Noyes Point, R. I. (Walcott, U. S. Geol. Surv.)

Again, if the coast be sinking, new deposits of coarse material may be made on older ones. In this way also great thicknesses of sedim

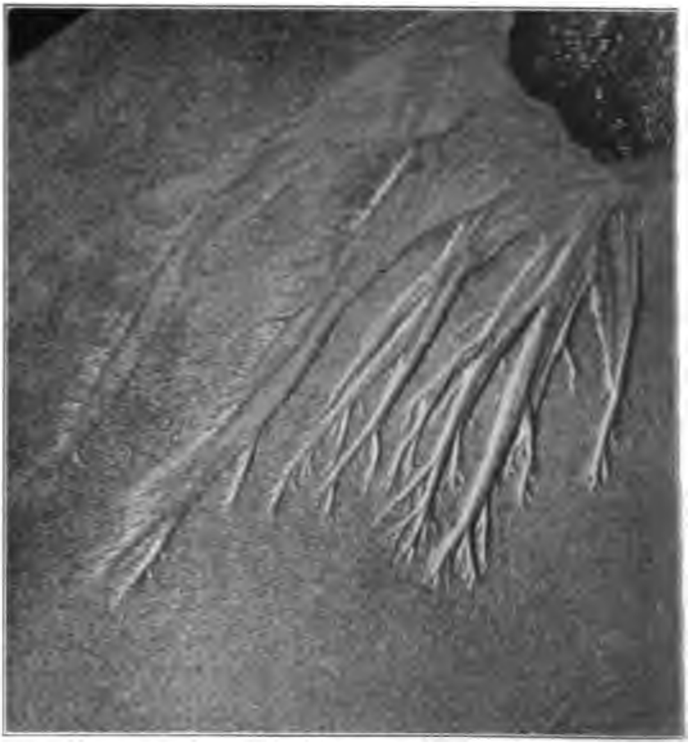

FIG. 326.—Rill-marks. Same locality as 325. (Walcott.)

may be accumulated, all parts of which were deposited in shallow w The great thickness of some of the conglomerate beds of the past how far this process may go.

As a rule, no definite line marks the seaward terminus of the detritus, since coarse material is carried farther out when the run high (and the undertow is strong) than when they are feeble calm weather, therefore, fine sediment may be deposited where had been laid down in the preceding storm, only to be covered by other deposits of a different character. Thus gravel grades sand, with more or less overlapping or interwedging, and sand off into silt in the same way. This is diagrammatically illustrate Fig. 323.

Characteristics of shallow-water deposits.—Clastic sediments laid down in shallow water have several distinctive characteristics. While they are, in the aggregate, coarse, they are characterized by frequent variations in coarseness. The surfaces of successive beds are likely to be ripple- and rill-marked (Figs. 324, 325, 326), and cross-bedding (Fig. 327) is of common occurrence. Clayey sediments accumulated between high and low water are often sun-cracked (Fig. 328), and the tracks of land animals are sometimes preserved on their surfaces. Shallow-water deposits often contain fossils of organisms which live in waters of slight depth. These characteristics are sufficient to differentiate sedimentary formations made in shallow water from those made in deep water, even after they have been converted into solid rock and after the rock has emerged from the sea. Many of these characteristics are, however, shared by deposits made by streams on the land. Subaërial and lacustrine sediments are usually distinguishable from those made in the sea by their fossils, and sometimes by their distribution.

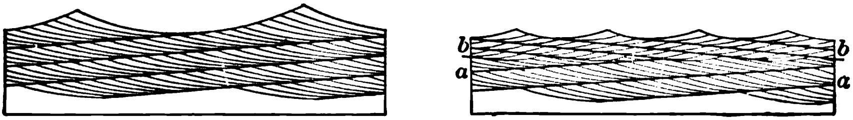

FIG. 327.—Cross-bedding. (Gilbert.)

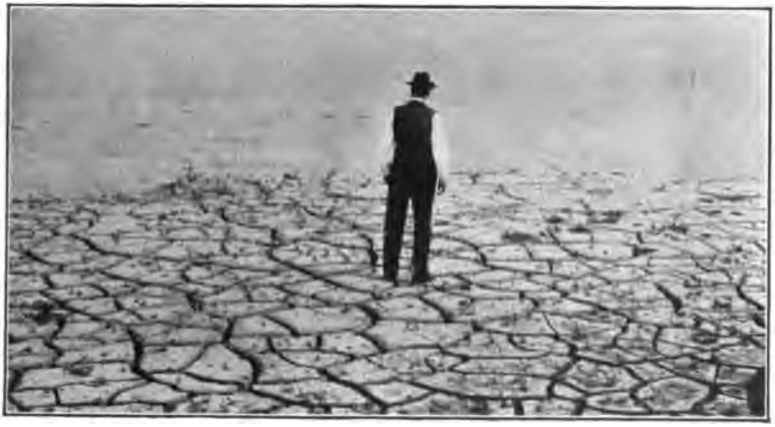

FIG. 328.—Sun-cracks. These cracks were on the mud-flats of the Missouri a few miles above Kansas City, but the sun-cracks on shore-deposits are not essentially different. (Calvin.)

Topography of shallow-water deposits.—The shallow-water deposits have, on the whole, a rather plane surface, though there are some notable departures from flatness. The steep slopes of the delta fronts and of wave-built terraces have already been spoken of. Barriers often shut in depressions, and the disposition of the material deposited is sometimes uneven, owing to shore and tidal currents. The result is that the surface of the shallow-water deposits is often affected by low elevations and by shallow depressions. The elevations and depressions may be elongate, circular, or irregular in form. These general facts are shown in Figs. 319, 320, and 329. This topography is sometimes preserved on newly emerged lands, as at various points on the Coastal Plain of the United States.

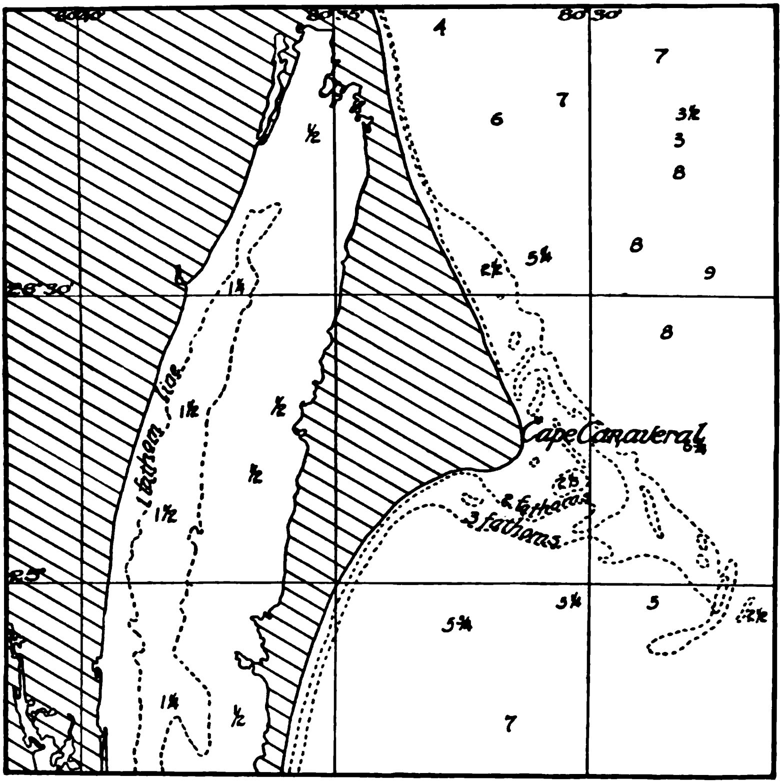

FIG. 329.—Irregularities of topography of shallow-water deposits. The depths of the water are shown in fathoms. (Chart of C. and G. Surv.)

Chemical and organic deposits.—There is no sharp line of distinction between the deposits usually classed as chemical and those regarded as organic. The latter are chemical in the broader sense of the term, but as they are immediately associated with life and are dependent upon it, it is a matter of practical convenience to separate them. Aside from the organic deposits, the chemical deposits made in shallow sea-water embrace (1) those due to reactions between constituents so brought together that new and insoluble compounds are formed and precipitated, and (2) those due to evaporation. The points of saturation for the various substances dissolved in sea-water are reached at different stages, and hence they are deposited more or less in succession.

The chemical deposits made in the shallow water of the sea, or in shallow bodies of water isolated from the sea, are chiefly simple precipitates resulting from evaporation; but new combinations are sometimes made in the process of concentration and precipitation. All substances in solution are necessarily precipitated on complete evaporation, but since the sea-water is in general far from saturation, so far as all its leading salts are concerned, only a few are thrown down in quantity sufficient to have geological importance where evaporation is incomplete. The leading deposits are lime carbonate ($CaCO_3$), lime sulphate (gypsum, $CaSO_4,2H_2O$), common salt (rock-salt, $NaCl$), and the magnesium salts, usually the chlorides and sulphates, which are later changed to carbonates. In investigations on Mediterranean water [1] which had an initial density of 1.02, no deposit took place until concentration by evaporation had brought the water to a specific gravity of 1.05. Between this density and that of 1.13, lime carbonate and some iron oxide were deposited. Between 1.13 and 1.22, lime sulphate was the most abundant precipitate, while between 1.22 and 1.31, 95% of the deposit was common salt. With still further concentration, the remaining substances in solution, especially the magnesium salts, were thrown down.

While there is somewhat more than ten times as much lime sulphate as lime carbonate in the ocean (p. 324), the deposits of the carbonate (including the organic) have been very much greater than those of the sulphate. This is due partly to the fact that the sulphate is much more soluble in natural waters than the carbonate. Rivers bring much more carbonate than sulphate to the sea, so that the point of saturation for the sulphate would normally be reached much later than that of the carbonate. The more important fact, however, is that marine plants and animals use lime carbonate freely for skeletal and housing purposes. It is held by some that they get their lime from the sulphate, but if so they convert it into carbonate before it takes the form of shells, coral, etc., the sulphuric acid set free in the process reproducing, directly or indirectly, more sulphate. The secretion of lime carbonate by organisms is not dependent on the saturation of the water, but may be carried on when the amount in solution is very small.

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[1] Usiglio. Encyclopædia Britannica. Article on Salt.

There can be little doubt that the chief deposits of lime carbonate have been and are being made through the agency of plants and animals in the form of shells, coral, bones, teeth, and other devices for supporting, stiffening, housing, protecting, and arming themselves; but while it is agreed that the larger part of the lime carbonate deposited in the open sea is of organic origin, it is equally clear that in closed seas subject to concentration from evaporation, simple precipitation takes place freely. There is some difference of opinion as to the importance of these two classes of deposits, past and present. The debated point is whether simple precipitation takes place in any appreciable degree under the usual oceanic conditions. There is much more evidence of solution by sea-water than of precipitation from it. The ocean appears to be under-saturated with lime carbonate on the whole, though it is still possible that deposition may take place in favorable situations, as, for example, where the very calcareous waters of rivers are spread out in thin sheets on the surface of the heavier salt water, and thus exposed to exceptional evaporation, or where there is very exceptional agitation and aëration.[1]

Gypsum appears to be deposited in quantity only in the closed basins of arid regions where concentration reaches an advanced state.

Since normal sea-water is far from saturation with common salt, the latter is precipitated only in lagoons, closed seas, or other situations favorable to great concentration. This is usually achieved only in notably arid regions, and in basins that receive little or no drainage from the land.

Deposits of salt usually, therefore, signify highly arid conditions, and where they occur over wide ranges in latitude and longitude, as in certain periods of the past, unusual aridity is inferred. Where confined to limited areas, their climatic significance is less, for topographic conditions may determine local aridity. The total area where salt is now being precipitated is small, though on the whole the present is probably to be regarded as a rather arid period of the earth's history. On the other hand, ancient deposits of salt preserved in the sedimentary strata show that the area of salt deposition has been much more considerable than now at one time and another in the earth's history. The salt and gypsum deposits of the past seem, therefore, to tell an interesting tale of the climates of the past.

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[1] Willis. Jour. of Geol., Vol. I, p. 500, where the evidences for deposition are fully set forth.

The magnesium salts are among the last to be thrown down as the sea-water is evaporated, and they most commonly take the form of sulphates and chlorides. They often form double salts with potassium, a relatively small and soluble constituent of sea-water. In the artificial evaporation of salt water to obtain common salt, the process is usually stopped before the saturation-point for the magnesium salts is reached, and the residue, the "mother-liquor," or "bittern," is drawn off to prevent these "bitter" salts from mixing with the common salt. The magnesium salts are among the last to be precipitated, not only because they are readily soluble, but because their quantity is small; yet in the original rock from which all the sea-salts came, there is at least as much magnesium as sodium, while in the sea there is about five times as much sodium as magnesium. Just what becomes of the remaining magnesium is not yet well understood. It has a notable disposition to form double salts with some other constituent, as noted above. In the earlier marine strata, dolomite, that is, limestone composed partly or wholly of the double carbonate of lime and magnesia, $(CaMg)CO_3$, abounds. This appears to have been formed by a gradual substitution of molecules of magnesium for those of calcium, but just how and when and why it was done has not been fully worked out. It appears to be a case where the saline matter of the sea made its contribution to the sedimentary deposits by chemical reaction upon them, rather than by precipitation because of saturation.

The relatively small amount of potash in the sea-water is probably due to its disposition to remain united with the clays and earths of the mantle rock and of the shaley deposits.

To some extent the salts in solution act directly on the earthy matter brought down into the sea by rivers, but where sedimentation is rapid, as it often is in shallow water, this action is limited and obscure. In the main, the ocean-waters protect the sediments from weathering and similar changes, except as organic matter buried with them induces change.

While the lime deposits are by far the greatest of the chemical and organic deposits of the sea, plants and animals also secrete notable quantities of silica. Silica deposits of organic origin are relatively much more important in the deep sea than in shallow water, and will be mentioned in that connection.

Limestone.—Something concerning the origin of limestone has already been given in the preceding paragraphs, but because of the importance of this formation, it may be added by way of summary that shallow seas free, or nearly free, from terrigenous sediment, and abounding in lime-secreting life, furnish the conditions for nearly pure deposits of limestone, and that most of the limestone within the areas of the present continents appears to have originated under such conditions. The common notion that limestone is normally a deep-water formation is a serious error. Although limestones are formed in deep as well as in shallow waters, by far the more important classes of lime-secreting organisms are photo-bathic, i.e. are limited to the depths to which light penetrates. In the shallow waters, these plants and animals are in part free and in part attached. Within the areas of deep water they are free and at the surface, and their remains drop to the bottom, if not sooner dissolved. But few forms live on the deep, dark, cold bottoms of abysmal depths. Clear waters, free from abundant terrigenous sediments and abounding in lime-secreting life, rather than deep waters, are, therefore, the most favorable conditions for the origin of limestone.

The purely chemical deposits of limestone are probably all of shallow-water origin. Once made, they are subject to solution, re-deposition. and other mutations like other deposits. As a result, they often lose many of their original characteristics, but enough usually remain to tell the story of their origin.

Deep-sea Deposits.

Contrasted with shallow-water deposits.—The deep-sea deposits cover the ocean-bottom below the 100-fathom line. Their area is considerably more than half the earth's surface. The characteristic deposits are muds, organic oozes, and clays, which in their physical characteristics are remarkably uniform. In regions of floating ice, greater diversity is introduced from the varied nature of the materials which the ice transports, but gravels and sands, comparable to those of shallow water, are rarely found. "Tides, currents, and waves produce some mechanical effects at the upper limits of the deep-sea region, but on the whole there is an absence of the phenomena of erosion, and mechanical action would appear to be absent except in the case of submarine eruptions. The depth is too great for sunlight to penetrate, and vegetable life is limited to the upper zone. Animal life is present in the same zone and on the bottom, but absent or nearly so in the middle depths. The temperature (at the bottom) is below 40° Fahr. throughout the larger part of the area, and if subject to variation with latitude or change of season, these changes affect only the depths immediately beyond the 100-fathom line. Throughout the whole region there is a very uniform set of conditions. In the shallow-water and littoral zones, owing to the rapid accumulation and the mechanical effects of transportation and erosion, the effects of chemical modification are not very apparent in the deposits; but in deep-sea deposits, in consequence of the less rapid rate of accumulation, absence of transport, the nature and small size of the particles, many evident chemical reactions have taken place, resulting in the formation *in situ* of glauconite, phosphatic and manganese nodules, zeolites, and other secondary products."[1] With increasing depth and distance from the shore, the character of the deposits undergoes a change. There is less and less material derived directly from the land, and more "amorphous matter arising from the ultimate decomposition of minerals and rocks, and accompanied, in all moderate depths, by an increase [relative] of the remains of pelagic organisms. We thus pass insensibly from those deep-sea deposits of a terrestrial origin, which we call 'terrigenous,' to those deep-sea deposits denominated 'pelagic,' in which the remains of calcareous and siliceous organisms, clays and other substances of secondary origin play the principal rôle."[1]

The following table[2] shows the relations of the various groups of marine deposits.

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[1] Murray, loc. cit.

[2] Ibid., p. 186.

1. Deep-sea deposits beyond 100 fathoms	Red clay Radiolarian ooze Diatom ooze Globigerina ooze Pteropod ooze	I. Pelagic deposits formed in deep water removed from land.
	Blue mud Red mud Green mud Volcanic mud Coral mud	II. Terrigenous deposits formed in deep and shallow water, mostly close to land.
2. Shallow-water deposits between low-water mark and 100 fathoms	Sands, gravels, muds, etc.	
3. Littoral deposits between high- and low-water marks	Sands, gravels, muds, etc.	

Sources.—The pelgaic deposits are made up in part of materials of organic origin, and in part of materials of inorganic origin. The inorganic materials may be of mechanical or chemical origin. Mechanical pelagic deposits originate in various ways. They may come (1) from the land by the ordinary processes of gradation, (2) from volcanic vents, or (3) from extra-terrestrial sources. Chemical deposits may be formed (1) *in situ* by the chemical interaction of substances in the sea-water on materials of organic and inorganic origin, and (2) by direct precipitation from the sea-water.

Mechanical inorganic deposits.—The terrigenous materials which reach the deep sea are, as a rule, only the finest products of land decay, and are carried out by movements of water or by the winds. They are not commonly recognized in the dredgings more than 200 miles from the shore, but opposite the mouths of great rivers they extend much farther,—1000 miles in the case of the Amazon. They are especially abundant on the slopes of the continental shelves. Here occur the *blue*, *green*, and *red muds*, with which are associated volcanic and coral muds. The color of these various muds is dependent in part on the changes which they have undergone since their deposition. The green muds usually contain enough glauconite to give them their color, and are most commonly found off bold coasts where sedimentation is not rapid. The blue muds indicate lack of oxidation, or perhaps deoxidation. Red muds are not common, though they have been found in some situations. In general, these deposits are analogous to certain shales, marls, etc., found within the continents.

Though coarse materials derived from the land are occasionally found in the deep-sea deposits, their presence must be looked upon as in some sense accidental. Occasional pebbles, or even bowlders, are carried out into the ocean entangled in the roots of floating trees. Within limits, too, icebergs have carried out land débris, though it is probable that transportation by this means has been exaggerated. The amount which icebergs might carry, if fully loaded, is far greater than the amount which they do carry.

Of the identifiable inorganic materials in the deep sea, the most abundant are of volcanic origin, and among these the most common is pumice, which is frequently so light that it floats readily until it becomes water-logged. Pieces of pumice brought up by the Challenger and thoroughly dried were found to float for months in sea-water before settling even through the depth of water contained in the vessel in which the experiment was performed.[1] The next most abundant substance of volcanic origin in pelagic deposits is *volcanic glass*. This ranges from pieces of the size of a walnut down to the smallest fragments, which often serve as centers for concretions. *Lapilli* (cinders) and *volcanic ash* also are abundant in parts of the deep sea. The distribution of these volcanic products is essentially universal, though by no means uniform. Some of them are probably from submarine volcanoes.

The study of the deep-sea deposits has revealed the presence of many nodules and grains which are believed to be of extra-terrestrial origin. Many of them are magnetic.[2] The dust of countless meteors which enter the atmosphere daily settles on land and sea alike, and enters into the sediment of the bottom of the latter. It is probably no more abundant in deep water than in shallow, but it is relatively more important, since other sedimentation is more meager. The number of meteorites which enter the atmosphere daily has been estimated at from 15,000,000 to 20,000,000.[3] If on the average the meteorites weigh ten grains each, probably a rather high estimate, the total amount of extra-terrestrial matter reaching the earth yearly would be 5,000 to 7,000 tons, and something like three-fourths of this must, on the average, fall into the sea. But even at this rate it would take some fifty billion years to cover the sea-bottom with a layer one foot in thickness.

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[1] Murray, loc. cit., p. 295.

[2] Challenger Report, Deep Sea Deposits, p. 327.

[3] Young's Astronomy, p. 472.

Organic constituents of pelagic deposits.—With increasing distance from shores, and especially with increasing depth of water, terrigenous deposits become less and less abundant, and sediments derived from pelagic life increase in relative importance. Beyond the upper part of the outer slopes of the continental shelves, the pelagic deposits are largely made up of shells and skeletons of marine organisms which live in the surface-waters. Pelagic molluscs, foraminifera, and algæ secrete shells of lime carbonate, while diatoms and radiolarians secrete shells of silica. When the organisms die, they sink to the bottom with their shells, and these mineral matters of organic origin are mingled with the volcanic products which are universal over the sea-floor. Pelagic deposits of organic origin are named according to their characteristic constituents. Thus there are pteropod oozes, globigerina oozes, diatom oozes, radiolarian oozes, etc.[1] It is not to be understood that these oozes are made up exclusively of the shells which give them their names. Diatom ooze is an ooze in which diatom shells are abundant, not an ooze made up wholly of diatom shells; and globigerina ooze is an ooze in which globigerina shells are abundant, though in many cases they do not make up even the bulk of the matter. While samples of these various oozes might be selected which are thoroughly distinct from one another, there are all gradations between them, since pelagic life does not recognize boundary-lines.

It is a significant fact that with increasing depth the proportion of lime carbonate in the ooze decreases. Thus in tropical regions remote from land where the depths are less than 600 fathoms, the carbonate of lime of the shells of pelagic organisms may constitute 80% or 90% of the deposit. With the same surface conditions, but with increasing depth, the percentage of lime carbonate decreases, until at 2000 fathoms it is less than 60%; at 2400 fathoms, 30%, and at 2600 fathoms, 10%. Beyond this depth there are usually no more than traces of carbonate of lime. The data at hand show that the percentage of lime carbonate falls off below 2200 fathoms more rapidly than at lesser depths.

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[1] Murray. Scottish Geog. Mag., Vol. XV, p. 511. An excellent summary of deep-sea deposits.

When the percentage of lime carbonate becomes very low, the calcareous oozes grade off into the *red clay* with which the sea-floor below 2400 to 2600 fathoms is covered.

Chemical deposits.—The chemical deposits of the deep sea are chiefly the alteration products of sediments which reach the sea-bottom by mechanical means. All sediment deposited in the sea undergoes more or less chemical change, but it is only when the change is very considerable that the product is referred to this class. Where sedimentation is rapid and the sediment coarse, the chemical change is relatively slight; but where the sedimentation is slow and the sediment fine, the chemical change is relatively great; for the longer exposure to the sea-water and the greater proportion of surface exposed to attack, both favor change. Both the area and the mass of sea-bottom sediment radically changed in this way are large, but most of the deposit does not correspond to any formation known on the land.

The red clay already referred to belongs to this class of deposits. Its origin has been the subject of much discussion. It contains much volcanic débris, various concretions, bones of mammals, zeolitic crystals, and extra-terrestrial spherules, and doubtless the insoluble products of the shells of pelagic life; but it is still a mooted question how far the clay itself is the product of decomposed shells, and how far the altered product of pulverized pumice, volcanic ash, dust, etc. Pelagic life does not seem to be less abundant at the surface where the water is deep than where it is shallow, and it would appear that the shells must sink in such situations as elsewhere. If the lime carbonate of globigerina ooze be removed by dilute acid, the inorganic residue is similar to the red clay in the ocean-bottom. This suggests that owing to the more complete solution in the very deep water, the lime carbonate of the shells has been dissolved, leaving the red clay as a residuum. The more complete solution at the bottom might be the result either of the greater pressure, or of a greater percentage of CO_2 in the water due to emanations from the sea-floor, or to both; but the suddenness of the transition from oozes to red clay, with increasing depth, does not seem to be fully explained by these assumptions. The study of the dredgings has inclined the students of these materials to the conclusion that volcanic materials, rather than shells, are the principal source of the red clay.[1]

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[1] Murray, Challenger Report on Deep Sea Deposits, p. 337 et seq., and Buchanan Proc. Roy. Soc. Edin., Vol. XVIII, 1892, pp. 17–39.

The volcanic materials are thought to have accumulated slowly and to have been long exposed to the action of sea-water. The various nodules and crystals in the clay are believed to be secondary products, the materials for which were derived from the decomposition of the same materials. Eolian dust may be a notable constituent of the red clay.

Various specific products of chemical change may be briefly referred to. The decomposition of certain mineral particles, such as feldspar, gives rise to *kaolin*, and kaolin is a very considerable constituent of most of the clayey deposits of the ocean-bottom. The kaolinization of feldspar may take place both on land and in the sea. *Manganiferous* deposits are widespread in the ocean-bottom, occurring both as coatings on grains of mechanical sediments, shells, etc., and as concretions ranging in sizes from minute particles to nodules an inch or more in diameter. The concretions are sometimes approximately spheroidal, but often botryoidal. These manganiferous nodules are believed to have arisen from the decay of fragments of volcanic rocks. In their decay, the manganese and iron are believed to have been first changed to carbonates, and subsequently to oxides. After manganese oxide, iron oxide and silica are by far the most abundant constituents, but many other substances enter into their composition in minor quantities.

Another substance somewhat widely distributed in the sea-bed, though by no means universal, is *glauconite*, a complex silicate of alumina, iron, potassium, etc. Glauconite is, on the whole, most abundant along the edges of the continental shelves, though it is by no means universal in this position. It is not commonly found in deep water, nor very near the shore, but approximately at the "mud-line." The glauconite grains begin to form, as a rule, in tiny shells, chiefly the shells of foraminifera. After filling the shell, the shell itself may disappear, while the glauconite goes on accumulating around the core already formed, until the grain attains considerable size. Glauconite is believed to be an alteration product of certain sorts of mechanical sediment, the change being effected under the influence of the decaying organic matter in the shells.[1] It does not occur where sedimentation is rapid, and its formation appears to be favored by considerable changes of temperature. Glauconite deposits occur on the land and are com-

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[1] Challenger Report on Deep Sea Deposits, pp. 385–391. See also Jour. of Geol., Vol. II, pp. 167–172.

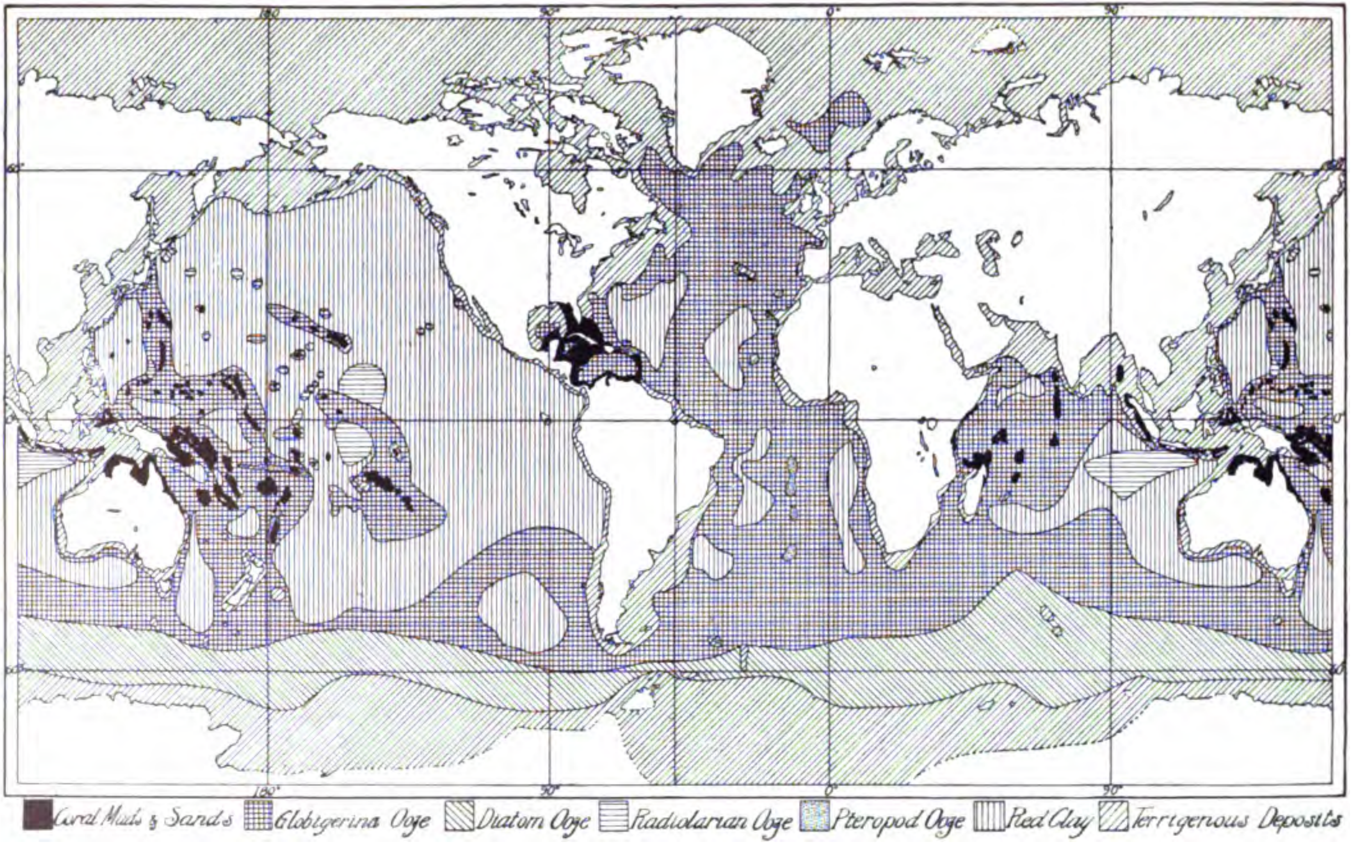

FIG. 330.—Distribution of various sorts of deep-sea deposits. (Murray. Challenger Reports.)

monly known as *green sand marl.* Glauconite also occurs sparingly in many other sedimentary rocks.

Another substance which is somewhat widespread in the ocean-bottom is *phosphate of lime,* which occurs in various sorts of oozes, in the manganiferous nodules, in glauconite, and in independent nodules. Like the grains of glauconite, the grains of phosphate of lime appear to have started as concretions in shells, and to be the result of the reaction of organic matter on the contents of sea-water. The immediate source of the lime phosphate in the water appears to have been the shells or bones of the numerous animals living in the sea.

Secondary minerals made from the constituents of volcanic matter which has been decomposed occur not uncommonly in the bottom of the sea. These minerals belong to the general class of *zeolites,* phillipsite being the most abundant. Their distribution is somewhat wide, but their quantity is slight.

Unfortunately, knowledge of the deep-sea deposits is limited to their superficial layers. Soundings do not usually penetrate more than a few inches, or at most a foot or two.

Unlike shallow-water deposits, those of the really deep sea seem to find no correlatives in the known rock formations of the land.

LAKES.

Most of the phenomena of the ocean are repeated on a smaller scale in lakes. The waves of lakes and their attendant undertows and littoral currents are governed by the same laws and do the same sort of work as the corresponding movements of the ocean. Tides are absent, or insignificant, but slight changes of level, known as *seiches,*[1] have been observed in many lakes. They are probably caused by sudden changes in atmospheric pressure. While they are generally very slight, they frequently amount to as much as a foot, and occasionally to several feet. The seiches are oscillatory movements, and their period is influenced by the length and depth of the lake. They have been studied most carefully in Switzerland. Currents corresponding to those of the ocean are slight or wanting in lakes, but since most lakes have inlets and outlets, their waters are in constant movement toward the latter. In most cases this movement is too slow to be readily noted, or to do effective work either in corrasion or transportation. The work of the ice, on the other hand, is relatively more important in lakes than in the sea.

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[1] Forel, Compte Rendu, 1875, 1876, 1878, 1879, and P. Du Bois, 1891. Also Forel's Lac Leman.

Changes taking place in lakes.—The processes in operation in lakes are easily observed and readily understood. (1) The waves wear the shores, and the material thus derived is transported, assorted, and deposited as in the sea, and all the topographic forms resulting from erosion or deposition along the seacoast are reproduced on their appropriate scale in lakes. (2) Streams bear their burden of gravel, sand, and mud into lakes and leave it there. (3) The winds blow dust and sand into the lakes, and in some places pile the sand up into dunes along the shores. (4) Animals of various sorts live in the lakes, and their shells and bones give rise to deposits comparable to the animal deposits in the sea. (5) Abundant plants grow in the shallow water about the borders of many ponds and lakes, and as they die, their substance accumulates on the bottom. (6) At the outlet the water is constantly lowering its channel. The lowering of the outlet is often slow, especially if the rock be coherent, for the outflowing water is usually clear, and therefore inefficient in corrasive work. These six processes are essentially universal, and all conspire against the perpetuity of the lakes. (7) In lakes where the temperature is low enough for ice to be formed, it crowds on the shores and develops phenomena peculiar to itself. The ice of the sea may work in similar ways, but its work is restricted to high latitudes. (8) In lakes in arid regions, deposits are often made by precipitation from solution. The first five and the last of these processes are filling the basins of the lakes. As the sediment is deposited, a corresponding volume of water is displaced, and, if there be outlets, forced out of the basins; the sixth process is equally antagnostic to the lakes, while the seventh has little influence on their permanence. Given time enough, these processes must bring the history of any lake to an end. The lowering of the outlet will alone accomplish this result if the bottom of the basin is above base-level. Many lakes have already become extinct, either through the filling or draining of their basins, or through both combined. The antagonism of rivers and lakes long ago led to the epigram "Rivers are the mortal enemies of lakes." True as this statement is, it does not follow that lakes will ever cease to exist, for the causes which produce new lakes may be in operation contemporaneously with those which bring lakes now in existence to an end.

Lacustrine deposits.—The beds of sediment deposited in lakes are similar in kind, in structure, and in disposition to beds of sediment laid down in the sea, but river-borne sediment is more commonly concentrated into deltas, since waves and shore-currents are less effective. Even the limestone of the sea has its correlative in some lakes. Some of it was made of the shells of fresh-water animals which throve where the inwash of terrigenous sediment was slight, some of it from the calcareous secretions of plants,[1] and some of it was precipitated from solution.[2] Salt and iron-ore [3] deposits are also sometimes made in lakes.

Extinct lakes.—The former existence of lakes where none now exist may be known in various ways. If the lake basin was filled, its former area is a flat, the beds of which bear evidence, in their composition, their structure, and often in their fossil contents, of their origin in standing water. Such a flat is commonly so situated topographically that the basin would be reproduced if the lacustrine deposits were removed. To this general rule there might be exceptions, as where a glacier formed one side of the basin when it was filled. If the lake was destroyed by the reduction of its outlet, or by the removal of some other barrier, such as glacier ice, or by desiccation, shore phenomena, such as beaches, spits, etc., may be found. In time such evidences are destroyed by subaërial erosion, so that they are most distinct soon after the lake becomes extinct.

Many lakes, some of them large [4] and many of them small, are known to have become extinct, while many others are now in their last stages, namely, marshes. Many others have been greatly reduced in size. Such reductions are often obvious where deltas are built into lakes. Thus the delta built by the Rhone into Lake Geneva is several miles in length, and has been lengthened nearly two miles since the time of the Roman occupation. The end of Seneca (N. Y.) lake has been crowded northward some two miles by deposition at its head. Similar changes have taken place and are now in progress in many other lakes.

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[1] C. A. Davis, Journ. of Geol., Vol. VIII, pp. 485–97, and 498–503, and Vol. IX, pp. 491–506.

[2] Russell, Lake Lahontan, Mono. XI, U. S. Geol. Surv., Chap. V; also Third Ann. Rept., pp. 211–221. Gilbert, Lake Bonneville, Mono. I, U. S. Geol. Surv., p. 167.

[3] Stapff, Zeit. deut. geol. Gesell., Vol. XVIII, pp. 86–173.

[4] Upham, Lake Agassiz, Mono. XXV, U. S. Geol. Surv.; Salisbury and Kümmel, Lake Passaic, Rept. of the State Geologist of N. J., 1893, and Jour. of Geol., Vol. III, pp. 533–560; Gilbert, Lake Bonneville, Mono. I, U. S. Geol. Surv.; Russell, Lake Lahontan, Mono. XI, U. S. Geol. Surv.; and Mono Lake, Eighth Ann. Rept., U. S. Geol. Surv., Pt. I.

Lake ice.[1]—Since fresh water is densest at 39° Fahr., ice does not commonly form on the surface until the temperature from top to bottom is reduced to this point. Cooled below this temperature, the surface-water fails to sink, and with sufficient reduction freezes. If the lake be small, and especially if it be shallow, it is likely to freeze over completely in any region where the temperature is notably below the freezing-point for fresh water for any considerable period of time. It is under these circumstances that the ice becomes most effective.

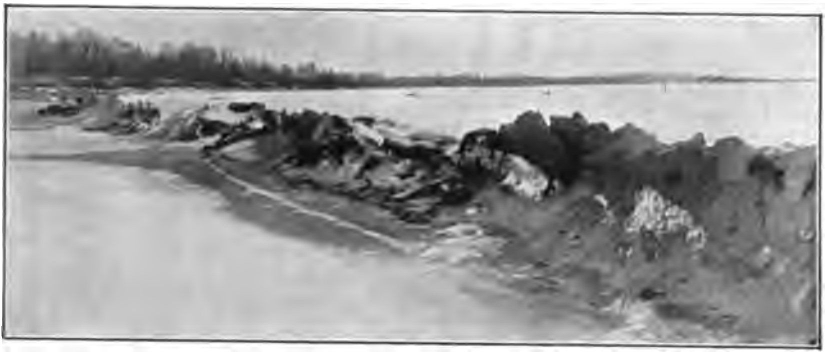

FIG. 331.—Ice crowding upon low shore. Clear Lake, Ia. (Calvin.)

Suppose a lake in temperate latitudes, where the range of temperature is considerable, to be frozen over when the temperature is 20° Fahr. If now the temperature be suddenly lowered to −10°, and such change of temperature is not uncommon in the northern part of the United States, the ice contracts notably. In contracting, it either pulls away from the shores or cracks. If the former, the water from which the ice is withdrawn quickly freezes; if the latter, water rises in the cracks and freezes there. In either case, the ice-cover of the lake is again complete. If the temperature now rises to 20° the ice expands. The cover is now too large for the lake, and it must either crowd up on the shores (Fig. 331) or arch up (wrinkle) elsewhere. It follows the one course or the other, or both, according to the resistance offered by the shore.

If the water near the shore is very shallow, the ice freezes to the sand, gravel, and bowlders at the bottom. If the adjacent land is low, the ice in expanding may shove up over it, carrying the débris frozen in its bottom. It may even push up loose gravel and sand in front of its edge if they be present on the shore. Where bowlders are frozen to the bottom of the ice, the shore-ward thrust in expanding has the effect of shifting them in the same direction, and even of lifting them a little above the normal water-level. This constant process of concentrating bowlders at the shore-line gives rise to the "walled" lakes, which are not uncommon in the northern part of the United States. The "wall" does not commonly extend entirely around a lake, though it exists at various points on the shores of many lakes. In making the walls, the ice shoved up by winds, especially in the spring when the ice is breaking up, cooperates.

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[1] Gilbert, Lake Bonneville, Mono. I, U. S. Geol. Surv., p. 71, and Topographic Features of Lake Shores, Fifth Ann. Rept. U. S. Geol. Surv., p. 109.

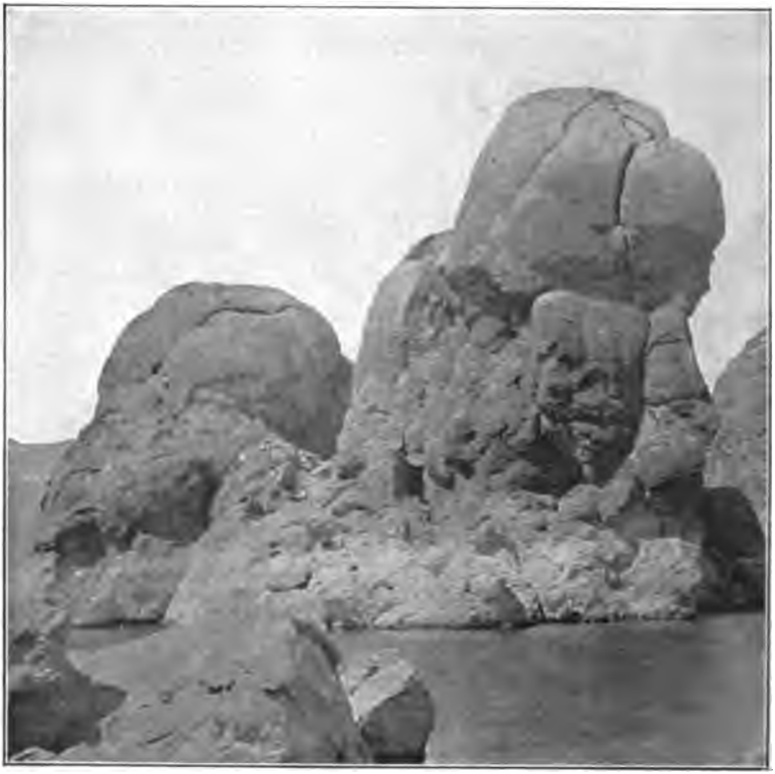

FIG. 333.—Calcareous tufa domes. Pyramid Lake, Nev. (Russell.)

If the lake be bordered by a low marsh, the ice and frozen earth of the latter are really continuous with the ice of the lake, and the push of the latter sometimes arches up the former into distinct anticlines, the frozen part only being involved in the deformation. A succession of colder and less cold periods may give rise to a succession of such anticlines.[1] If the shore be steep and of non-resistant material, the crowding of the ice produces different but not less striking results. Where the thrust of the ice is against a low cliff of yielding material, such as clay, it disturbs all above the shore-line. Where the cliff is sufficiently resistant, it withstands the push of the ice, and the ice itself is warped and broken.

Saline lakes.—A few lakes, especially in arid or semi-arid regions, are salt, and others are "bitter." Beside sodium chloride, salt lakes usually contain magnesium chloride, and magnesium and calcium sulphates. "Bitter" lakes usually contain much sodium carbonate, as well as some sodium chloride and sulphate, and sometimes borax. The degrees of saltness and bitterness vary from freshness on the one hand to saturation on the other. The water of the Caspian Sea (lake) contains, on the average, less salt than that of the sea; that of Great Salt Lake contains about 18%; that of the Dead Sea, about 24%; and that of Lake Van (eastern Turkestan), the densest body of water known, about 33%. See accompanying table.

Many salt lakes, such as the Dead Sea and Great Salt Lake, are descended from fresh-water ancestors, while others, like the Caspian and Aral Seas, are probably isolated portions of the ocean. Lakes of the former class have usually become salt through a decrease in the humidity of the region where they occur. The water begins to be salt when the aridity is such that evaporation from the lake exceeds its inflow. In this case the inflowing waters bring in small amounts of saline and alkaline matter, which is concentrated as evaporation takes place. The concentration may go on until the point of saturation is reached, or until chemical reactions cause precipitation. In general the least soluble minerals are precipitated first. Thus gypsum begins to be deposited from sea-water when 37% of it has been evaporated; but the saturation-point for salt is not reached until 93% of the water has been evaporated (see p. 375). The relations in lakes are similar, and gypsum deposits often underlie those of salt. Deposits of salt and other mine matters once in solution are making in some salt lakes at the pres time, and considerable formations of the same sort have been so ma in the past. Buried beneath sediments of other sorts, beds of comm salt or of other precipitates are preserved for ages. Lime carbon has been precipitated in quantity from some extinct lakes (Fig. 333).

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[1] Buckley. Wis. Acad. of Sci., Vol. XIII, Pt. I, 1900. A study of ice ramparts formed about the shores of Lake Mendota, Wis., in 1898–99.

The lakes which originate by the isolation of portions of the sea a salt at the outset. If inflow exceeds evaporation, they become fresh and may ultimately become fresh; otherwise they remain salt. If eva oration exceeds inflow they diminish in size and their waters beco more and more salt or bitter.

Indirect effects of lakes.—Lakes tend to modify the climate of t region where they occur, both by increasing its humidity and by decre ing its range of temperature. They act as reservoirs for surface-wate and so tend to restrain floods and to promote regularity of stream flo They purify the waters which enter them by allowing their sedimen to settle, and so influence the work and the life of the waters below.

Composition of lake-waters.—The accompanying table[1] shows the com position of various inclosed lake-waters, and gives some idea of the wi range, both in kind and quantity, of the mineral matter held in soluti by them. It is to be noted that the table shows the composition of t waters of exceptional, rather than common, lakes. The waters of fre lakes do not depart widely from those of rivers (p. 107).

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[1] Copied from Russell's Lake Lahontan, Mono. XI, U. S. Geol. Surv.

OF THE

ARTS PER 1000

ad Sea, ale, surface	ake	Great Salt	Urmiah Lake	Van Lake	Aral Sea
1.0216	170		1.155	....	....
0, 1864	....	1869	....	....	....
....	....	O. D. Allen.	Hitchcock.......	ancourtois	....
Geological loration of d Sea, p. 278	xpe- reat .419	U. S. Geo Expl. 40 1877, V p. 435	Lartet Geological Exploration o Dead Sea, p.28	chof's Chemical Geology, Vol. I, p. 94	Roth Chemical Geology, p. 465
.885		49.69	74.890	8.502 [4]	2.4512
.474		2.40	......	.246	.0584
.....		......	......	......	.0022
2.150		.25	.529	......	.4581
4.197		3.78	2.914	.157 [4]	.5965
.....		Trace	......	......	......
Trace		.....	......	Trace [6]	.0008
7.628		83.94	119.496	5.693	3.8386
.167		Trace	......	......	.0029
race		.....	......	5.267 [4]	.0918
.202		9.85	7.671	2.555	3.3368
.....		.....	......	......	.0011
.....		.....	......	......	Trace
.....		Trac	......	......	......
.006		.....	......	.180	.0032
race		.....	......	......	......
.....		.....	......	......	......
race		.....	......	......	Trace
race		.....	......	......	Trace
5.709		149 9	205.500	22.600	10.8416

[1] Less .04254 Average of two a [7] Carbonic acid b

CHAPTER VII.

THE ORIGIN AND DESCENT OF ROCKS.

It has been the current opinion that the earth was once in a molten state, and thence cooled to a solid condition, and hence that all the primitive rocks were igneous. Even those who think that the earth may never have passed through a molten state agree that the oldest known rocks are either true igneous rocks, or rocks of very similar nature. A molten magma may, therefore, be taken as the mother state of the rocks. Starting with this conception, the natural order of events suggests the inquiries (1) how rocks are formed from molten magmas, (2) what natures they assume, (3) how other rocks are derived from them, (4) how still other rocks are derived from these derivatives, and so on. To answer these inquiries is to trace out the generations of rocks and learn the general history of rock-formation.

(1) The process by which igneous rocks are formed from lavas is actually taking place in existing volcanoes. As these are widely scattered over the face of the earth, the material poured out by them represents different parts of the interior and varies in nature accordingly. This affords the means of studying the differences that arise from differences of material. This is a radical consideration, for variations in composition give rise to the most fundamental distinctions between rocks, though by no means the only ones. Rocks which have the same composition often differ greatly in texture or structure, owing to the varying conditions under which they were formed. In the solidification of rocks from the molten state, the rate of cooling causes many differences. A means of studying this is afforded by the various lava flows that are now being poured out on the surface under different conditions; but a more important means is afforded by extinct volcanoes, especially by those which have been deeply cut open by erosion. In certain very ancient volcanoes, not only have the solidified lava streams of the surface been cut across by erosion, but the lava that remained in the crater, or in the neck that led up from below, is laid bare for inspection. Exposures of even more profound nature have been made by the great

disruptions which the outer part of the crust has suffered. In certain tracts there have been profound fractures, and the formations on one side of these have settled down and on the other side have been pushed up (faulted), so as to expose parts that were once much below the surface. Sometimes also the crust has been folded and crumpled, and the wrinkles thus formed have afterwards been worn away or cut open by deep valleys, and rocks that were once deeply buried have been laid bare. By the revelations made in these and other ways, it has been learned that at various times in the history of the earth molten matter has been thrust into fissures or intruded between layers of the crust and cooled there, without coming to the surface. Sometimes the lava appears to have forced its way into the rocks, and sometimes to have lifted the upper beds and formed great subterranean layers or tumor-like aggregates, called bathyliths and laccoliths (Fig. 334). Such intruded bodies of molten rock, solidifying under the varying conditions of such subterranean situations, are a fruitful source of instruction respecting the influence of varying rates and modes of cooling, as well as of other attendant conditions.

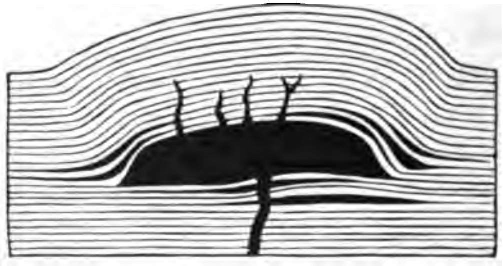

FIG. 334.—Diagram of a laccolith. (Gilbert.)

It will thus be readily seen that the rate of cooling of the various molten rocks must have differed very greatly. In the portions poured out upon the surface there were sometimes narrow streams and thin sheets, giving large exposure in proportion to the mass (Fig. 335), and sometimes thick flows and deep pondings in basins and choked valleys, giving massive bodies with relatively small surface exposure. There were explosions of the lava into minute particles with almost instantaneous cooling, and there were eruptions beneath the sea the peculiar effects of which are rather matters of inference than of positive knowledge. In the portions underground there were insinuations into thin fissures, on the one hand, and in-thrustings of thick bodies, on the other. Some intrusions entered the upper part of the crust where the rocks were cold and wet, and some were thrust into the deeper portions where the rocks were warmer and less penetrated by water. Sometimes the lava rose rapidly and was little cooled in passage, sometimes slowly with more cooling en route, and sometimes there were long halts between eruptions, with much opportunity to cool. An almost infinite variety of conditions is thus presented, and with it a rich field for the study of the modes of solidification.

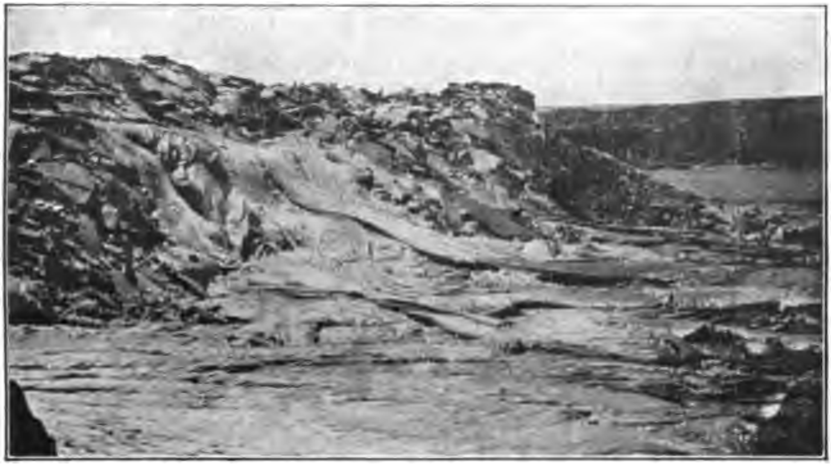

Fig. 335.—Fresh lava flow, with large surface exposure. Holemaumau, Hawaii. (Libbey.)

In the underground intrusions the additional factor of high pressure was also present, and this is the third important condition in determining the nature of igneous rocks.

The three factors, *composition*, *rate of cooling*, and *degree of pressure*, require special consideration.

Composition of Igneous Rocks.

All or nearly all the chemical elements known on the earth are found in greater or less amounts in igneous rocks, and in a broad sense are constituents of them. If there are any exceptions, they are most likely to be found in the rarer elements in the atmosphere. Oxygen, nitrogen, hydrogen, aqueous vapor, and carbonic acid, which make the mass of the present atmosphere, are all found in lavas and in th cooled products. Probably all the rarer elements also occur in igneous rocks. Helium is known to be given forth by springs.

Leading elements.—But although nearly or quite all the know chemical elements enter into the igneous rocks, only a few of them ar abundant. These are regarded as normal or essential constitue while the rarer substances are regarded as incidental. By combin a large number of the most trustworthy analyses of rocks of all sor F. W. Clarke[1] has estimated the relative amounts of the more abund elements in the crust of the earth with the following result:

Element.	Symbol.	Percent. in the Solid Cr
Oxygen	(O)	47.02
Silicon	(Si)	28.06
Aluminum	(Al)	8.16
Iron	(Fe)	4.64
Calcium	(Ca)	3.50
Magnesium	(Mg)	2.62
Sodium	(Na)	2.63
Potassium	(K)	2.32
Titanium	(Ti)	.41
Hydrogen	(H)	.17
Carbon	(C)	.12
Phosphorus	(P)	.09
Manganese	(Mn)	.07
Sulphur	(S)	.07
Barium	(Ba)	.05
Strontium	(Sr)	.02
Chromium	(Cr)	.01
Nickel	(Ni)	.01
Lithium	(Li)	.01
Chlorine	(Cl)	.01
Fluorine	(Fl)	.01
		100.00

It will be seen that only eight of the elements hold a high ra quantity. Many that are of the utmost importance in the hist the earth and the affairs of men are low in the list, or do not even ap in it at all, because their quantity is too small to be estimated in centages. The precious metals, and even some of the more com metals, as lead, zinc, and copper, are too scarce to form an appre percentage.

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[1] Analyses of Rocks, Bull. 168, U. S. Geol. Surv., 1900, p. 15.

Union of elements.—In a general study of the igneous rocks we may for the present neglect all but the first eight of these elements. Out of these elements spring various chemical combinations, and out of these combinations spring the various minerals, and out of the combinations of minerals come the various rocks. The union of oxygen with the other seven elements may be taken as a fundamental step in this series of combinations. The result is the following oxides: Silica (SiO_2), alumina (Al_2O_3), ferrous, ferric, and magnetic oxide (FeO, Fe_2O_3, and Fe_3O_4), magnesia (MgO), calcium oxide or lime (CaO), soda (Na_2O), and potash (K_2O). The oxygen sometimes unites in proportions different from those here given, but such exceptions may be neglected in a general study. We thus have nine leading oxides. Of these, silica acts as an acid, or more strictly according to the newer chemical view, as an acid anhydride. All the rest, except the magnetic oxide of iron, and sometimes the oxide of aluminum, act as basic oxides.

In the older chemical philosophy these oxides were supposed to combine by the simple union of an acid oxide with a basic oxide, and to remain as oxide joined to oxide; thus silica (SiO_2) and lime (CaO) formed silicate of lime (CaO,SiO_2). The symbols express the idea better than the words. This method is used in the older geological works and in some of the later. But in the newer chemical doctrine, the oxides are not believed to remain so distinct after their union, and the symbols are written $CaSiO_3$, and the compound is named calcium silicate. According to the modern doctrine of solution, some of the calcium, silicon, and oxygen may exist as free ions in molten rock. The precise way in which the elements are related to each other in these compounds can scarcely be said to be known. For the general purposes of geology it is most convenient to think of these oxides as uniting in the simple fashion first named, and this involves no apparent geological error in general studies, since they are oxides when they enter the compound, and if the compound is decomposed they usually come forth again as oxides; but in closer studies more complex unions, attended by dissociations (ionization), must be recognized.

Formation of minerals.—As but one of the leading oxides that abound in an average magma plays the part of an acid, the silica, a very simple conception of the general nature of igneous rocks may be reached by noting that they are mostly silicates of the seven leading basic oxides —alumina, potash, soda, lime, magnesia, and the iron oxides. This general idea is a very useful one and represents a most important truth; but in its use we must not forget that there are many exceptions. Sulphur, phosphorus, chlorine, and other elements unite with the bases to form sulphates, sulphides, phosphates, phosphides, chlorides, etc. So also there are many minor bases that form silicates; and these minor bases unite with the minor acids to form many more or less rare minerals. Again, there are native metals in some igneous rocks. But altogether these hardly reach more than one or two percent. of the whole.

There are, however, two exceptions of more importance. In the molten magma the acid and basic elements are not always evenly matched. When there is an excess of silica, a portion remains free and takes the form of quartz (SiO_2). If there is an excess of the basic oxides, the weakest one is usually left out of the combination. This is commonly the iron oxide, which then usually takes the form of magnetite (Fe_3O_4). It is a singular fact that quartz often forms when there is no excess of silica, and magnetite when there is no excess of base. Quartz (free acid anhydride) and magnetite (free basic oxide) sometimes occur in the same rock. The explanation for this is yet to be found. These form rather important exceptions to the generalization that the igneous rocks are mostly made up of silicates, but, thus qualified, it expresses the essential truth, and has the merit of embodying the central chemical fact relative to these rocks.

Sources of complexity.—But here simplicity ends. As we pass on to the specific silicates that are formed, we encounter several sources of complexity. In the first place, the silica unites with the bases in different ratios and thus gives rise to unisilicates or orthosilicates (ratio of oxygen of bases to oxygen of silica, 1:1), subsilicates (ratio more than 1), bisilicates (ratio 1:2), trisilicates or polysilicates (ratio 1:3 or higher), and combinations of these. All the bases are not known to combine in all these ways, but many do in more than one of them. Still, if the silica were content to unite with each of the bases by itself alone, the results would remain comparatively simple; but instead of this it unites with two or more at the same time; and, more than that, it unites with them in varying amounts. The case would still remain measurably simple if these chemical compounds always crystallized out by themselves, each compound forming one mineral, and but one; but the different silicates have the confusing habit of crystallizing together in the same mineral. A crystal may thus sometimes be seen, under the microscope, to be made up of alternating layers of different silicates; e.g., a microscopic layer of an aluminum-calcium silicate may be overlain by a microscopic layer of an aluminum-sodium silicate, and the alternation may be repeated throughout the crystal, giving it a banded structure. There is reason to believe that this is true in many cases where the microscope fails to detect it, and that less symmetrical comminglings of silicates may take place. As such alternations or mixtures are not governed by any known mathematical law, as is the case in chemical compounds, there is no determinate limit to the number of combinations that may arise. As a matter of fact, new ones are still being discovered in the progress of research, and the total number that may ultimately be found can scarcely be prophesied.

As a result of all this fertility of combination, the total number of silicious minerals in igneous rocks is large. It is the function of the mineralogist to treat of these minerals as such. The geologist deals with them as constituents of the earth and as factors in its history. Only a few of them are so abundant as to require special individual notice in a general study of the earth. It may be remarked also that only a few of them can be identified by simple inspection as they occur in the rocks, partly because of the delicacy of the distinctions between many of them, and partly because of their minuteness and intricate intermixture. The resources of the polarizing microscope are necessary for safe determination in most cases. The student need not feel embarrassment or discouragement if he is often unable to recognize the constituents of the intimately crystalline rocks. Their determination has grown to be a profession by itself.

The leading minerals of igneous rocks.—Fortunately for the simplicity of geological study, a few minerals make up the great mass of the igneous rocks. These few are *quartz*, the *feldspathic minerals*, the *ferromagnesian minerals*, and the *iron oxides*. Quartz (silica, SiO_2) is the free acid already mentioned. The feldspathic and ferromagnesian minerals are the leading silicates of the earth's crust, and vastly surpass all others in abundance. The feldspathic group embraces minerals formed by silica in union with alumina, together with either potash, soda, or lime, or two or more of these together. The ferromagnesian group embraces minerals formed by the union of silica with iron, magnesia, and lime, together with more or less of the other basic oxides.

These statements are only true in a very general sense. Admixtures, replacements, and impurities are so frequent as to break down all sharp, simple definitions. The feldspathic minerals are normally light in color, ranging from white to red or gray. The ferromagnesian minerals are normally dark (commonly greenish) from the presence of iron, the great coloring element of rocks. But these color distinctions do not hold good in detail and cannot be much trusted as a means of identification.

The feldspathic minerals (p. 462) embrace the potash feldspars, *orthoclase* and *microcline;* the soda feldspar, *albite;* the lime feldspar, *anorthite;* and the mixed feldspars intermediate between albite and anorthite, viz., the soda-lime feldspar, *oligoclase,* the lime-soda feldspar, *andesine,* in which lime and soda are nearly equal, and the lime-soda feldspar, *labradorite,* in which the lime predominates; together with *leucite,* a potash silicate higher in alkali than orthoclase, and *nephelite,* a soda silicate higher in soda than albite. Leucite and nephelite are usually classified as *feldspathoids,* not as feldspars. It is to be understood that alumina is normally present in all these. Additional details respecting these minerals may be found in the reference list, p. 460.

Among the **ferromagnesian minerals** the most important are the pyroxenes, the amphiboles, and the biotite type of mica. Olivine is of subordinate importance. The pyroxenes (p. 465) and amphiboles (p. 460) have nearly the same chemical composition, but differ in crystallization and physical properties. *Hornblende* (an amphibole) has been melted, and on cooling under proper conditions found to take on the form of *augite* (a pyroxene). Pyroxene is sometimes altered into *uralite,* one of the amphiboles. The pyroxenes and amphiboles are the most abundant of the dark minerals in crystalline rocks. The leading members of the pyroxene group are *augite, diallage, hypersthene, enstatite,* and *soda pyroxene.* The chief members of the amphibole group are *hornblende* and the *soda amphiboles.* All are essentially silicates of magnesia and iron oxide, with or without the addition of lime, soda, and alumina. Details respecting these may be found in the reference list.

The two leading **micas** are the iron-magnesia mica, *biotite,* and the potash mica, *muscovite,* the familiar "isinglass" of the stove-door. Chemically, muscovite should go with the potash feldspars, but it is distinguished from them by its crystalline habit and physical properties. The biotite should go chemically with the pyroxenes and amphiboles, which it closely resembles except in its crystalline properties. Details respecting the micas may be found in the reference list, p. 464.

Two **iron oxides,** magnetite (Fe_3O_4) and hematite (Fe_2O_3) are widely disseminated in igneous rocks. They constitute the free bases already mentioned.

Summary of salient facts.—The salient facts are, therefore, (1) that out of the seventy-odd chemical elements in the earth, eight form the chief part of it; (2) that one of these elements uniting with the rest forms nine leading oxides; (3) that one of these oxides acts as an acid and the rest as bases; (4) that by their combination they form a series of silicates of which a few are easily chief; (5) that these silicates crystallize into a multitude of minerals of which again a few are chief; and (6) that these minerals are aggregated in various ways to form rocks. Possessed of these leading ideas, we are prepared to turn to the consideration of some of the conditions under which these combinations take place in the formation of rocks from molten magmas.

THE NATURE OF MOLTEN MAGMAS.

We easily fall into the habit of thinking of molten rock as we think of a molten metal, merely as a substance which has passed from the solid to the liquid condition because of high temperature. With the return of low temperature a molten metal returns to the solid state usually in the same molecular condition which it possessed before. The point of fusion and the point of solidification are the same and are rigidly fixed. If this were true of the constituents of a rock, a definite order for the solidification of the several minerals might be anticipated. As a matter of fact, the order is not the same under all conditions, and, what is especially significant, the order is far from being that in which the constituents would fuse or would solidify separately. For instance, in a granite composed chiefly of quartz, feldspar, and mica, the quartz is often the last to take form, although it is more infusible than the feldspar or the mica. This and other phenomena show that a molten magma is not to be viewed simply as a fused substance, but rather as a *solution* of one silicate in another, or as a solution of several silicates in one another mutually. The high temperature is to be regarded merely as a condition prerequisite to solution, or as the condition of fusion of some one constituent which then dissolves the others. If crystals of snow, sugar, and salt be mixed at a low temperature and compacted, the mass may be regarded as an artificial rock. On raising the temperature, all will pass into solution while the temperature is still somewhat below the melting-point of the snow, the most fusible, and while it is much below that of either the sugar or the salt. This particular case is instructive because the ice is not simply fused by temperature; the affinity of the salt plays a part. If the temperature were again lowered, the sugar and salt would not crystallize out at their fusing-points, but would remain in solution down to and even below the normal freezing-point of water; in other words, they would remain in solution until the water crystallized out and forced them to take the solid state. This holds good when the amounts of the sugar and salt are small relative to the water. If, on the contrary, their quantity is large relatively, crystallization will take place at higher temperatures and before the water crystallizes to ice. From this it appears that the salt and sugar might crystallize either before the water or after it, according to the degree of concentration. The behavior of mixtures of minerals in passing into and out of the molten condition appears to be quite analogous to this, and hence a great variety of results attend the process, dependent upon the number, the nature, and the relative quantities of the ingredients. The approved conception of the genesis of a rock from a molten magma (when ample time is given) is that one compound after another crystallizes out as the temperature falls and its point of *saturation* for each is reached, until the whole has been solidified. The modes of combination of the elements in the molten magma are not necessarily the same as those in the derivative crystals; indeed, the combinations doubtless change as the process proceeds; certain constituents being taken out, the remaining ones probably rearrange themselves.

Time required in crystallization.—The liquid magma of igneous rocks is essentially a fluid glass or slag. It is analogous to common glass, which is a silicate of potash, soda, or other base, except that usually common glass is relatively free from iron and other coloring substances, while these abound in the natural magmas and render them dark and more or less opaque; but the fundamental nature is the same, except that the natural lavas are usually mixtures of several silicates, while the artificial glasses consist of only one, or at most a few. Furnace slag is essentially an artificial lava.

When a lava is cooled quickly, the commingled silicates solidify in the diffused condition essentially as they were in the liquid; for there is no time for the silicate molecules of a like kind to come together, particle by particle, in regular systematic order, as required in crystallization. The essential feature of crystallization is this systematic arrangement of the molecules according to a definite plan, giving a specific crystal form, as a cube, a hexagonal prism, etc.

There are six (sometimes made seven) fundamental systems of crystallization, and a multitude of variations of special form in each system. The treatment of these forms belongs to mineralogy.

In a thick viscid liquid, this systematic arrangement of molecules into definite crystal forms takes place slowly, for the crystalline force in the silicates is far less energetic than that in water, which crystallizes into ice with much rapidity and with great force. Because of this slowness, the solidification of the lava may catch the process of crystallization at any stage. If the lava is cooled quickly, the result is a glass; if less quickly, part glass and part crystals; if slowly enough, all becomes crystalline. In general the slower the growth the larger the crystals. The solidification product may, therefore, range from a glass to a mass of crystals; i.e., it may be (1) wholly glass, (2) a glassy matrix with a few small crystals scattered through it, (3) a less abundant glassy matrix with more and larger crystals, (4) a mere remnant of glass in a mass of crystals, or (5) a mass of crystals with no glass.

Successive stages of crystallization.—Since eruptions take place intermittently, it is obvious that cooling of the lava may be in progress in its hidden reservoir during the quiescent intervals between eruptions. After a certain stage of partial crystallization has been reached during such time of quiet, a renewal of eruption may take place and the whole mass of lava be shifted into quite new conditions, and a second phase of solidification may be superposed on the one already started. The rock will then show two phases of crystallization: (1) large crystals of the kind or kinds most prone to develop in the given lava may have grown during the first long stage of slow subterranean cooling, while the greater part of the lava still remained liquid; and (2) small crystals or glass may have developed when the more rapid cooling under the new conditions took place. The result would be large crystals set in a matrix of small crystals or of glass, a combination styled *porphyritic*. In such cases the lava, in its later stages, carries the large crystals floating throughout its mass, and is not a simple liquid.

THE FRAGMENTAL PRODUCTS OF SUDDEN COOLING.

Pyroclastic rocks.—The extreme example of sudden cooling is presented when lavas are violently exploded into the air and solidify almost instantly. The resulting glassy particles or filaments, if small, constitute *volcanic ash.* The explosion appears to be due to steam and other gases which are held in the deeper lava under great pressure, but which, as they rise toward the surface of the lava where the pressure is relieved, expand with explosive violence. It is probably also due in part to progressive crystallization, which forces the gases out from the part that crystallizes and overcharges the rest. Sometimes the projected particles draw after themselves long *filaments* like the threads of spun glass, and sometimes while in the air they divide and draw apart, spinning a filament of viscid lava between them. A variety of this kind at the volcano of Kilauea in Hawaii is known as "Pele's hair." These light filaments drift with the wind and lodge on the lee side of the volcano, covering the surface "like mown grass" (Dana).

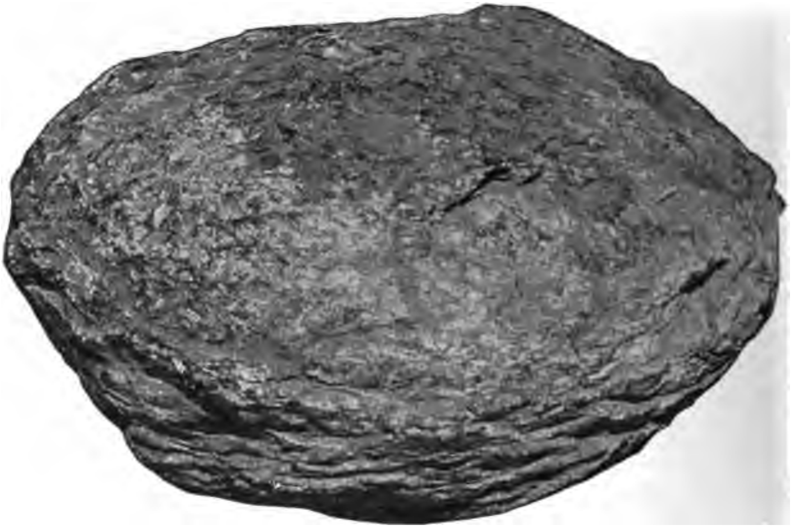

Fig. 336.—Volcanic bomb. About half natural size. (Photo. by Church.)

When the exploded fragments are coarser they fall about the volcanic vent and form the *tuffs* (*tufa*) of which most steep volcanic cones are chiefly built. In these larger fragments, crystals are not infrequently found, and the same is even true of the volcanic ash. These crystals are undoubtedly such as had already been formed in the lava before it exploded, and their formation, as suggested above, may have contributed to the explosion.

Fragments too large to be borne far away by the air, but still small, are known as *lapilli*, especially if they are somewhat rounded and gravel-like. A finer variety, of the nature of sand, much used in making Portland cement, is locally known as *puzzolana.*

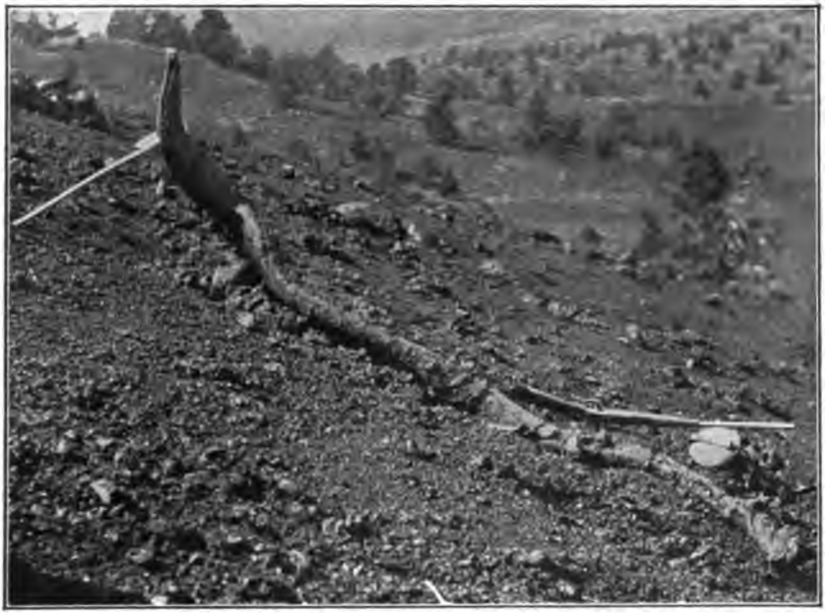

FIG. 337.—Volcanic bomb of unusual form, 13 feet long. Cinder Buttes, Idaho. (Russell, U. S. Geol. Surv.)

The rougher, irregular fragments of a clinker-like nature ejected by volcanoes are known as *scoriæ* or *cinders.* They are more or less distended by gas-bubbles and are hence light and pumiceous.

The larger masses of lava ejected into the air are often caused to rotate by the unequal force of the projection, or by the unequal friction of the air, and to assume spheroidal forms, the internal gases at the same time often expanding and rendering the mass vesicular. These rounded projectiles are known as *volcanic bombs* (Figs. 336 and 337). Balls of lava that have originated in rolling movements of the seething mass, or in other ways, are also styled bombs. Usage is not altogether harmonious or consistent in the application of the term.

The larger masses that are projected into the air are more or less vesicular from the expansion of included gases, as already noted, and so the fragmental products of volcanic action grade into the vesicular. The type of this class is *pumice*, in which the gas cavities make up by far the larger part of the volume of the whole mass, and the whole is reduced to the condition of a solidified froth or foam. So thin are the dividing films of glassy material in some cases that the whole is pure white, though the same material in solid mass would be dark. This solidified glassy froth is often lighter than water and floats freely on the sea until it becomes "water-logged" and sinks. Dredgings of the deep sea show that much pumice has accumulated there, and being far from the land has escaped burial by the sediments borne in by the rivers.

All of these fragmental rocks produced by volcanic action are known as *pyroclastic* (fire-fragmented) rocks, a general term of much convenience in distinguishing them from lava-flows, on the one hand, and from the fragmental rocks produced by air and water (ordinary clastics), on the other.

THE GLASSY ROCKS.

The solid glasses.—The quick cooling of lava-flows into solid glasses is chiefly dependent on their exposure at the surface. Hence it is often the case that the exterior of a lava-flow is glassy in greater or lesser degree, while the interior is more or less crystalline. Quick cooling is sometimes also due to the intrusion of the lava in thin sheets into fissures in cold rocks. When massive bodies of lavas penetrate solid rocks, the lava does not usually cool so fast as to prevent some degree of crystallization, and the crystallization may even become complete; but if the intruded lava sheet be very thin, the lava is liable to be cooled to a nearly perfect glass. The glassy condition is, therefore, subject to indefinite gradations. As a rule, the acid lavas are stiffer at the same temperature than the basic ones, and crystallize more slowly, so that acid glasses are more common than basic ones. The basic rocks usually crystallize pretty thoroughly, except on the immediate surface of the flows.

The first stages of crystallization.—The microscopic study of the volcanic glasses reveals great numbers of minute forms known as *crystallites*, *microlites*, *globulites*, etc., that appear to be first steps in crystallization, though many of them do not take definite geometrical shapes and some do not show the optical characters of crystals. There are minute globules (globulites), needles, and hair-like bodies (trichites) of more or less indeterminate nature, together with other forms that can be seen to be certainly the initial forms of well-known minerals.

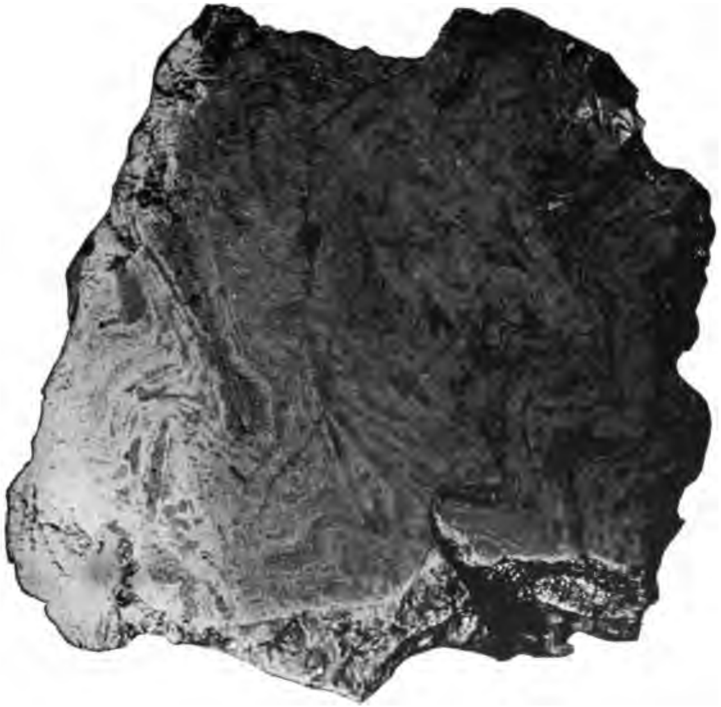

Fig. 338.—Flow structure in rhyolite. Nearly natural size. (Photo. by Church.)

The obsidians.—Of the compact glassy rocks, *obsidian* is the best type. It is essentially a natural glass, formed usually of acid silicates. It has the close texture, conchoidal fracture, and other qualities of glass. It is usually black, but sometimes red, brown, purple, bluish, or gray. While chiefly of glass, it usually contains more or less of the incipient crystals above described, showing that even here the first step in the gradation to the next or the crystalline stage has been taken. These incipient crystals sometimes become so abundant as to change the texture from the vitreous to the stony order. In some cases, the stony texture seems to have been developed in the obsidian after it was formed, the change being a part of a subsequent process of devitrification, but in other cases the crystals seem to be original. Besides these, there are often small globular bodies known as spherulites.

Varieties of glassy rock in which the embryo crystals are more numerous and the glassy texture less perfect, are known as *pitchstones*. The fresh surfaces of these have rather the aspect of pitch or resin than that of true glass; hence their name. Like the obsidians, they are usually dark, but they take on greenish, brownish, yellowish, and light-colored hues as well. Sometimes glassy rock fractures in small spheroidal forms like pearls, and is known as *perlite*. Basic glasses are relatively rare, and while usually included under the term obsidian, are sometimes given special names.

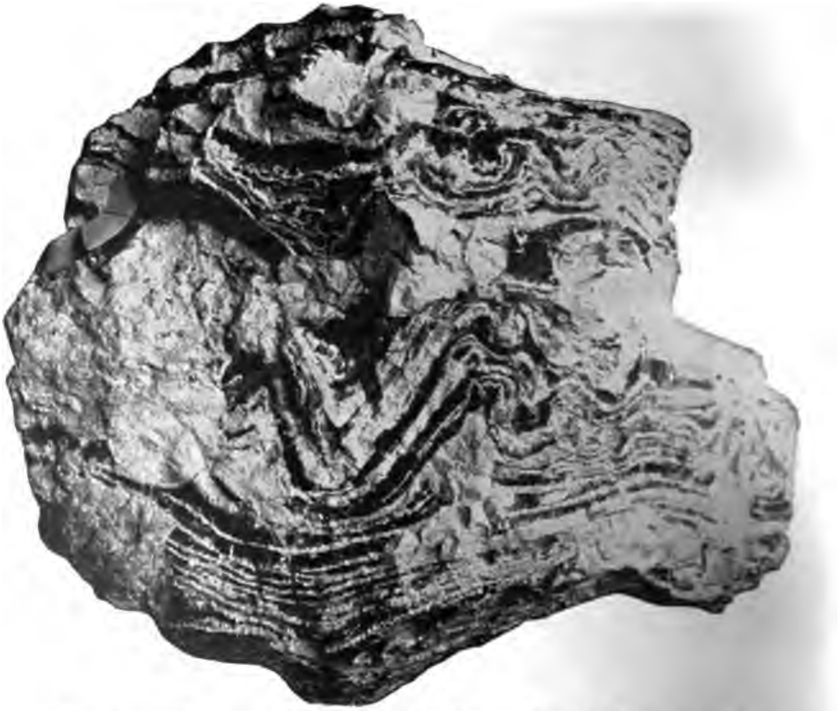

FIG. 339.—Flow structure in volcanic glass. About half natural size. (Photo. by Church.)

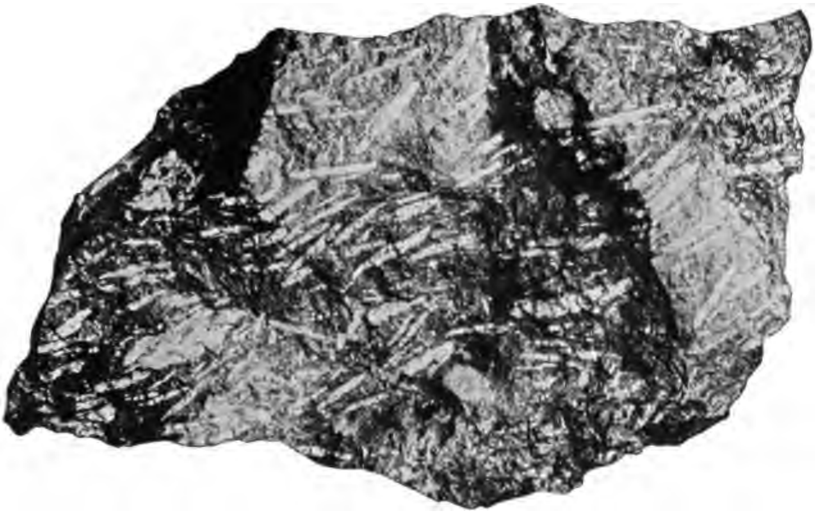

FIG. 340.—Flow structure in porphyry, shown by the position of the large crystals. About two-thirds natural size. (Photo. by Church.)

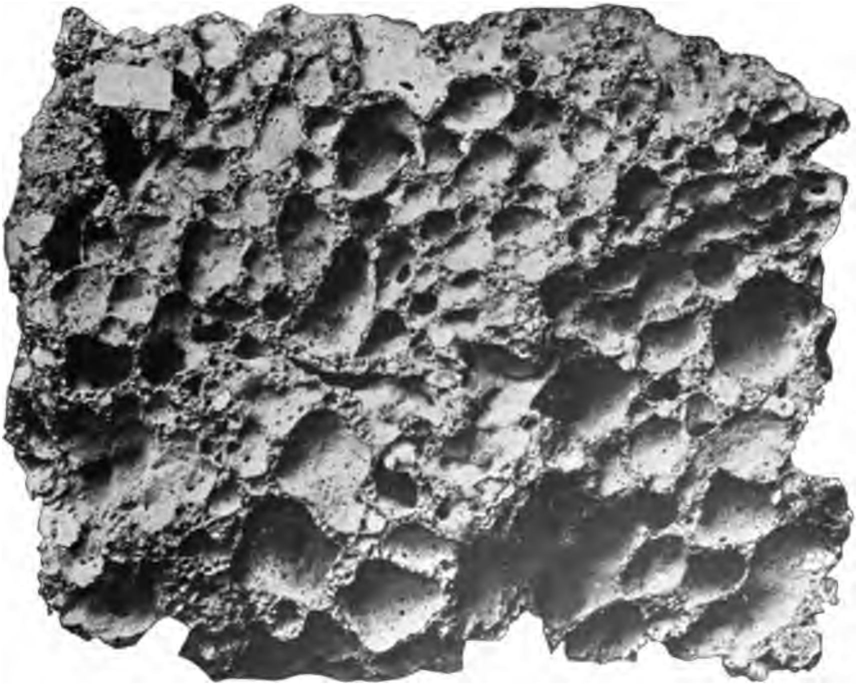

FIG. 341.—Scoriaceous texture. About four-fifths natural size. (Photo. by Church.)

SPECIAL STRUCTURES.

Flow structure.—Lavas that cool into glassy rocks frequently contain gas cavities, colored spots and variations of texture which, together with the hair-like embryo crystals, are drawn out into lines, streaks, and parallel belts by the flow of the viscous mass, giving rise to *rhyolitic* or flow structure (Figs. 338 and 339). Rocks in which this is the most pronounced feature were formerly known as rhyolites, though the term has drifted away from this original meaning and has been applied to a class of acidic rocks. The obsidians and pitchstones may be more or less rhyolitic under the microscope, though to the naked eye they may appear only as a glassy or resinous mass. The rhyolites generally have but an imperfect glassy texture, since the crystals and the cavities some-es make up a notable part of the mass, the glassy portion being ely more than a matrix in which the crystals, spherulites, and cavities

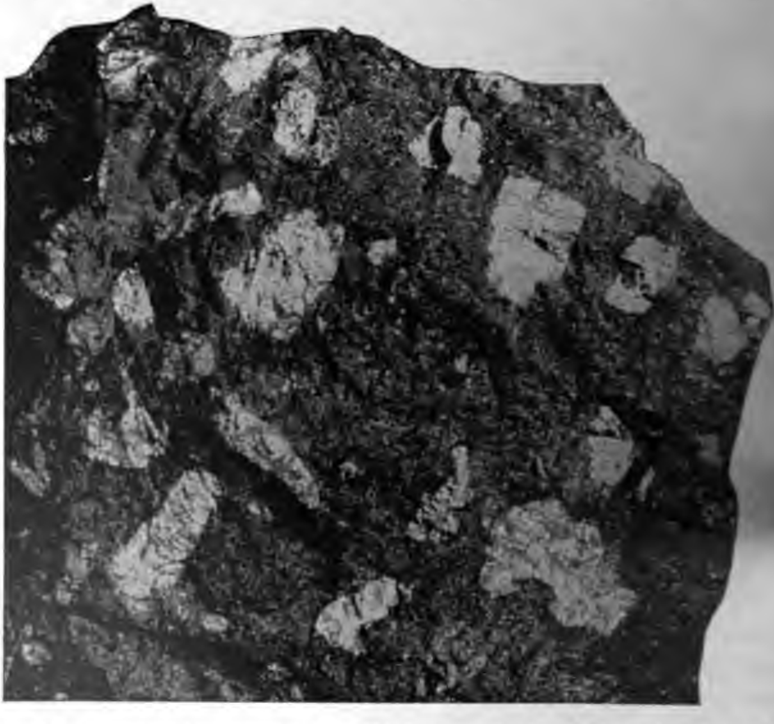

FIG. 342.—Porphyritic texture. Two-thirds natural size. (Photo. by Church.)

are carried. By an increase of the crystals in number and size, the rock passes by gradations into porphyry or phanerite.

Amygdaloids.—In lava-flows the included steam often collects in bubbles near the surface as the lava cools and forms a vesicular portion with a scoriaceous texture (Fig. 341). In its upper part, the vapor bubbles may be numerous, while below they become more and more scattered until they disappear. Similar bubbles are also often found near the bottom of a sheet of lava. This is perhaps due to the rolling under of the frontal surface of the lava-stream as it flows. Later, these cavities often become filled with minerals deposited from solution and the rock then becomes an amygdaloid, but this filling is a secondary action.

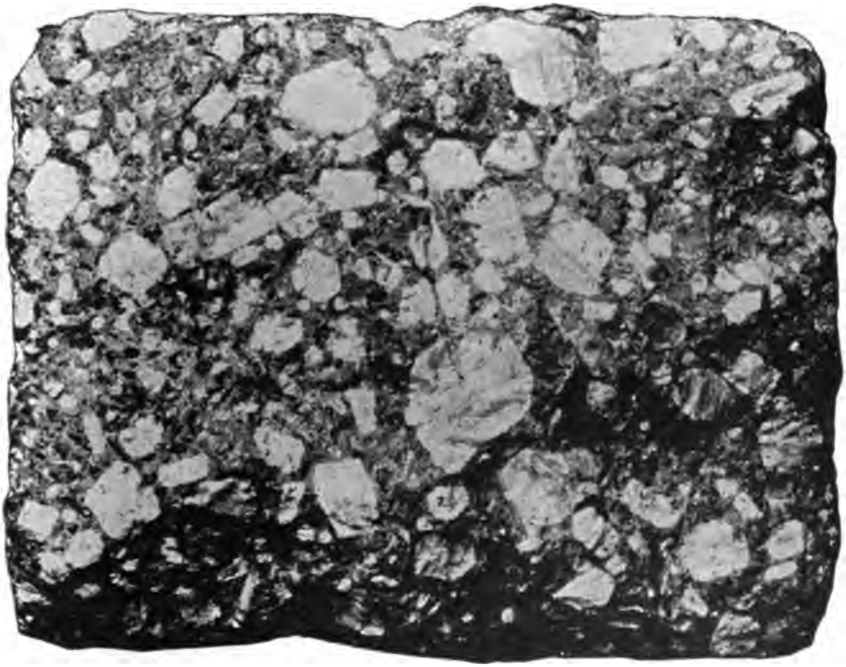

Fig. 343.—Porphyritic texture. Natural size. (Photo. by Church.)

THE PORPHYRITIC ROCKS.

When the conditions are such that after a part of the magma has formed distinct crystals floating in the remaining liquid lava, there is a change which causes the rest to solidify as a glass or as a mass of small crystals, the structure is known as *porphyritic*, and the rocks possessing it are called *porphyries*. This differentiation into distinct crystals set in a ground-mass of minute crystals or of glass often gives a mottled or variegated aspect to the rock, especially if the matrix of glass or minute crystals differs in color from the distinct crystals. This structure is much oftener developed in acidic rocks than in basic ones, because the latter crystallize more readily. The most common porphyritic crystals are feldspar and quartz, though they are by no means the only ones. The matrix is also usually felsitic or quartzose, but not necessarily so. The character is a structural one, and is not dependent upon any special chemical or mineralogical constitution. The distinct crystals are known as phenocrysts, and the varieties of porphyries are named from the characteristic phenocryst, e.g., quartzophyre (quartz-porphyry) if the conspicuous crystals are quartz, orthophyre if orthoclase, augitophyre if augite, etc. A convenient classification has recently been proposed[1] into (1) *leucophyre* (white porphyries), which have a light-colored ground-mass set with phenocrysts of any kind, and (2) *melaphyres* (black porphyries), which have a dark-colored ground-mass, with phenocrysts of any kind. While it is to be hoped this usage will prevail, it is to be noted that these terms, especially the latter, have been used in a different sense. (See reference list of rocks, p. 445.)

In many cases the ground-mass itself becomes minutely crystalline and the porphyritic aspect is due simply to large distinct crystals set in a mass of minute obscure ones. The rock is then really *holocrystalline*, but the term porphyry is applied to it. In other rocks the crystals of the ground-mass become more and more distinct, the porphyritic aspect gradually disappears, and there is a graduation into the next class.

THE PHANEROCRYSTALLINE ROCKS.

The phanerites.—When time enough is given for the cooling process the molten magma becomes completely crystalline. The holocrystalline rocks hence include a large series, ranging from the most acid to the most basic. In this class the differentiation of the rock material and the formation of distinct minerals reach a high stage, and as a natural result the varieties of rock are numerous. Taken as a group they are phanerites. If they are to be more particularly characterized, it is usually done on the basis of the minerals of which they are composed. The following are the leading types, beginning with those which are rich in silica and poor in basic oxides, and ending with those which are rich in basic oxides and poor in silica.

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[1] Quantitative Classification of Igneous Rocks, by Whitman Cross, Joseph P. Iddings, Louis V. Pirsson, and Henry S. Washington. 1903.

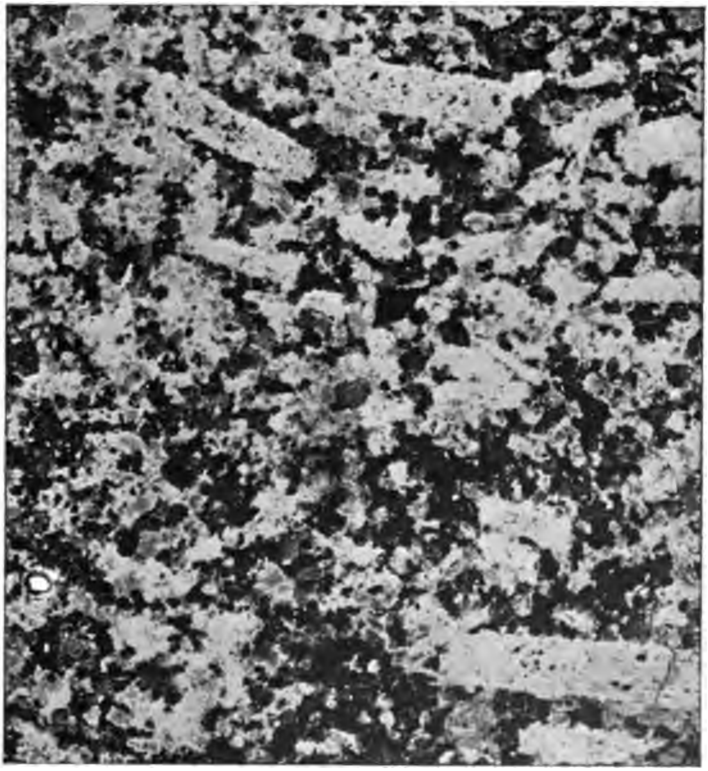

FIG. 344.—Granitic texture. About half natural size. (Photo. by Church.)

The granites.—The term granite was originally used to designate a granular, i.e., a distinctly crystalline, rock, and it is still popularly and properly so used. In scientific treatises it has usually been confined to a special aggregate of crystals of quartz, feldspar, and mica. It has recently been proposed to give it again a more general application, though not quite its original one, by including under it all holocrystalline rocks composed of dominant quartz and feldspar of any kind, with mica, hornblende, or other minerals in subordinate amount. In scientific literature as it now stands, granite consists of quartz, feldspar, and mica, the feldspar being of the alkali-potash or soda variety (orthoclase, microcline, or albite), and the mica, either muscovite or biotite. In the type form the crystals are distinct and sometimes large (Fig. 344). They are intimately mingled with one another, and in growing, interfered more or less with each other and so became interlocked. The granites are among the most common and easily recognized of the holocrystalline rocks. Their color is mainly dependent upon the feldspar, the red and pink varieties of the mineral giving rise to red granite, and the white varieties to gray granite.

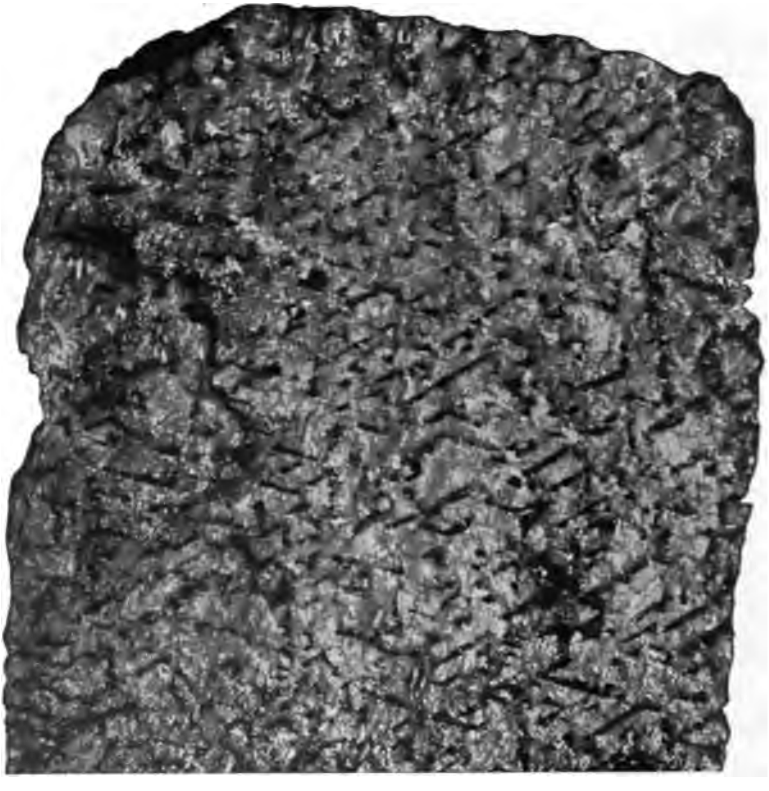

Fig. 345.—Graphic granitic (or pegmatitic) texture. Nearly natural size. (Photo. by Church.)

Very few granites conform strictly to the type. They vary by the addition and substitution of other minerals, and these sometimes become as prominent as the type minerals. The soda-lime feldspars sometimes take the place of the orthoclase, or accompany it; hornblende and other minerals take the place of the biotite, or occur with it; and so on. Whenever one of these replacing or accessory minerals is notable in quantity, its name is often prefixed, as hornblende-granite, oligoclase-granite, zircon-granite, etc. In this way the rock grades almost insensibly into the syenites, diorites, etc. Variations also arise from the absence of one of the three leading minerals. If mica is absent, the rock is termed an *aplite* (quartz and feldspar). If the feldspar is absent, it is called a *greisen* (quartz and mica). If quartz is absent, it is termed a *minette* (feldspar and mica). These varietal terms are neither universally nor always consistently used, and it is to be hoped they will be replaced by the systematic nomenclature recently proposed and outlined later (p. 451).

The granites were formed from a magma rich in silica, alumina, potash, and soda, but generally poor in lime, iron, and magnesia. Incidentally other substances were present. The alumina, potash, and other bases united with so much of the silica as was required to form the feldspars and micas, and the remaining silica crystallized into quartz.

Granite is normally a massive rock without foliation or banding. If it takes on these characters, it becomes a *gneiss*, and passes into the foliated or schistose class of rocks, to be discussed later. The texture of graphic granite (see pegmatite) is notably peculiar, due to the simultaneous crystallization of the quartz and feldspar (Fig. 345).

The syenites.—When the mica of a granite is replaced by hornblende, the rock is now commonly known as a *hornblende-granite*, but it was formerly called *syenite*, because found at Syene on the Nile. The term syenite is now applied to a rock consisting essentially of feldspar and hornblende or mica, but there is a complete gradation from the granites to the syenites. The magma of the syenites was richer in iron and magnesium than the typical granitic magma. The syenites also grade into other classes, as do the granites, and are named by similar prefixes, as augite-syenite, etc., and some of these varieties have special names. The syenites are red or gray, according to the color of the feldspar, and are usually darker than the granites. The texture of syenite is like that of granite. In the scheme of field names recently proposed, syenite is made to include all holocrystalline rocks composed mainly of feldspar of any kind, with subordinate amounts of mica, hornblende, pyroxene, and other minerals, but without a noticeable amount of quartz.

The diorites.—These embrace rocks which were crystallized from a magma still poorer in silica and the alkalies, and richer in the earthy bases. In composition they closely approach the ideal average rock, but usually fall a little below it in silica and the alkalies, and rise a little above it in the earthy bases. In current usage, diorite is defined as an intimate mixture of crystals of hornblende and a plagioclase feldspar. It differs from the syenite in having plagioclase feldspar instead of orthoclase. By substitutions and the addition of accessory minerals, the diorites graduate toward the granites and syenites on the one hand, as already noted, and into gabbros on the other.

In the scheme recently proposed, all holocrystalline rocks in which hornblende is dominant and feldspar subordinate are classed as diorites.

The gabbros.—The name gabbro was formerly applied to a coarse-grained basic rock consisting of labradorite and diallage, but the name has been gradually extended until it embraces a large group of rocks that have essentially the same composition as the dolerites mentioned below, but are coarser in crystallization, and the crystals do not embrace one another (i.e., are not ophitic). The principal minerals are plagioclase (normally labradorite) and pyroxene (normally diallage) with magnetite or ilmenite. They are usually dark, heavy rocks. The pearly luster of the cleavage faces of the diallage, when present, gives a peculiar sheen to a fresh surface of the rock. In the recently proposed field names, gabbro is made to include all phanerocrystalline rocks in which pyroxene predominates, attended by feldspar of any kind in subordinate quantity, with or without hornblende or mica.

The peridotites.—These stand at the basic end of the series, having been formed from a magma in which the silica was low (39–45 per cent), as were also the alumina, lime, and alkalies, but in which the magnesia was relatively very high, ranging from 35 to 48 per cent. The rock consists very largely of olivine associated with pyroxene, magnetite, and other very basic minerals. Little or no feldspar is present. The peridotites are much less abundant than the preceding classes and repre-nt a very distinctive phase of the magma in which the magnesia was atly concentrated.

Closely allied to the peridotites are rocks which are made up largely of a single basic mineral, as *augitite, pyroxenite, hornblendite,* rocks essentially formed of the minerals augite, pyroxene, and hornblende respectively. It will be noted that in these rocks the magma became quite simple in nature, just as at the acid end of the series certain rocks become comparatively simple from the concentration of the acid element, as in certain acidic granites, felsites, etc. (See pp. 523–524.)

The basalts.—The term basalt is used in a somewhat comprehensive way to embrace dark, compact, igneous rocks that appear to be nearly homogeneous, owing to the minuteness of the crystals, which are usually so small as to be identifiable only under the microscope. In some cases the crystals are scattered throughout a ground-mass after the porphyritic fashion. In some of these cases there is a true glassy base, and in such cases the rock does not strictly belong in the holocrystalline group. In the more typical cases the constituent minerals are very minutely crystallized and intimately intermixed. The leading minerals are plagioclase (usually labradorite or anorthite) and pyroxene (usually augite), with olivine and magnetite or ilmenite usually present. There is a considerable range in chemical nature, but the basalts are relatively poor in silica, usually also low in potash and soda, but rich in lime, magnesia, and the iron oxides. They are classed as basic and are sometimes highly so. The magmas of the basalts are especially fluid, and when poured forth upon the surface easily spread out in thin sheets. In cooling they are prone to take on a columnar or basaltic structure, the columns standing at right angles to the surfaces exposed to cooling. The columns are sometimes curved, owing to the peculiar attitude of the cooling surface. The columns of Giant's Causeway and Fingal's Cave are familiar examples.

The dolerites.—The basalts graduate insensibly into the dolerites; indeed the dolerites may be regarded simply as basalts of coarser crystallization. The minerals are evident to the eye and range up to medium size. The more abundant minerals are plagioclase feldspar (labradorite or anorthite), with one or more of the ferromagnesian minerals (augite, olivine, or biotite), and magnetite or ilmenite. In the growth of the minerals one crystal frequently embraces others, giving an ophitic structure. The dolerites have many varieties, due either to accessory minerals or to the development of some of the constituents more amply than the rest. The type may be said to consist of plagioclase and augite, the other minerals being regarded as accessories. Magnetite or ilmenite is almost universally present. The varieties are usually designated by prefixes, as olivine-dolerite, enstatite-dolerite, etc., but special names are also used for some of these.

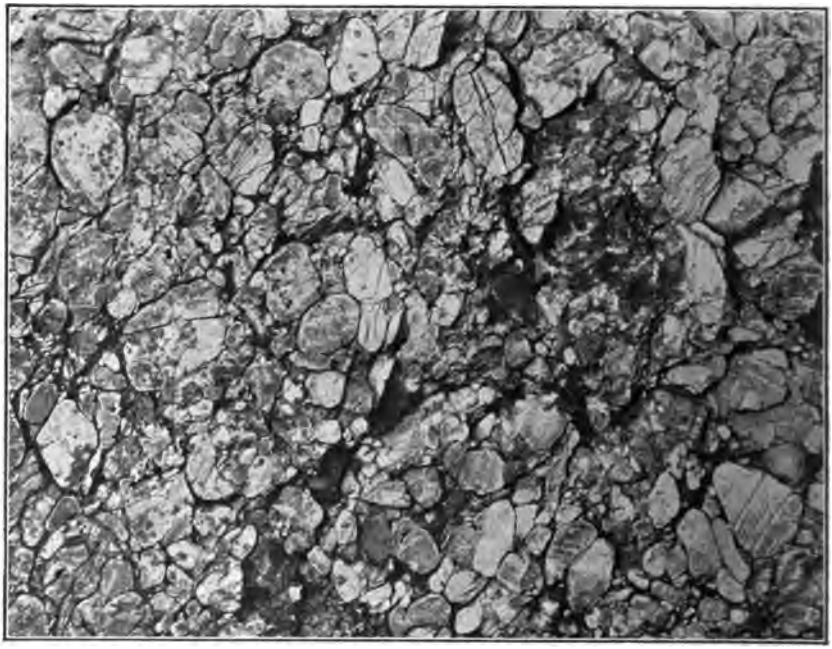

FIG. 346.—Conglomerate, Carboniferous series. Bancroft Place, Newport, R. I. (Walcott, U. S. Geol. Surv.)

The ancient dolerites have usually undergone internal changes and such rocks are often called *diabases*. While the use of the term has not been uniform, it accords with the better practice to regard the diabases simply as partially altered dolerites and basalts. In general, therefore, the diabases are but ancient dolerites.

General names.

The difficulty of distinguishing many of the foregoing rocks from each other by any means available in the field, owing to the minuteness of the crystals, and to the gradation of one type of rock into another, makes it desirable to employ certain general names which will correctly express the leading character of the rock without implying a knowledge of the precise mineral composition. A convenient term of this kind is *greenstone*, which merely indicates that the ferromagnesian minerals are prominent and usually give a greenish or dark cast to the rock. The

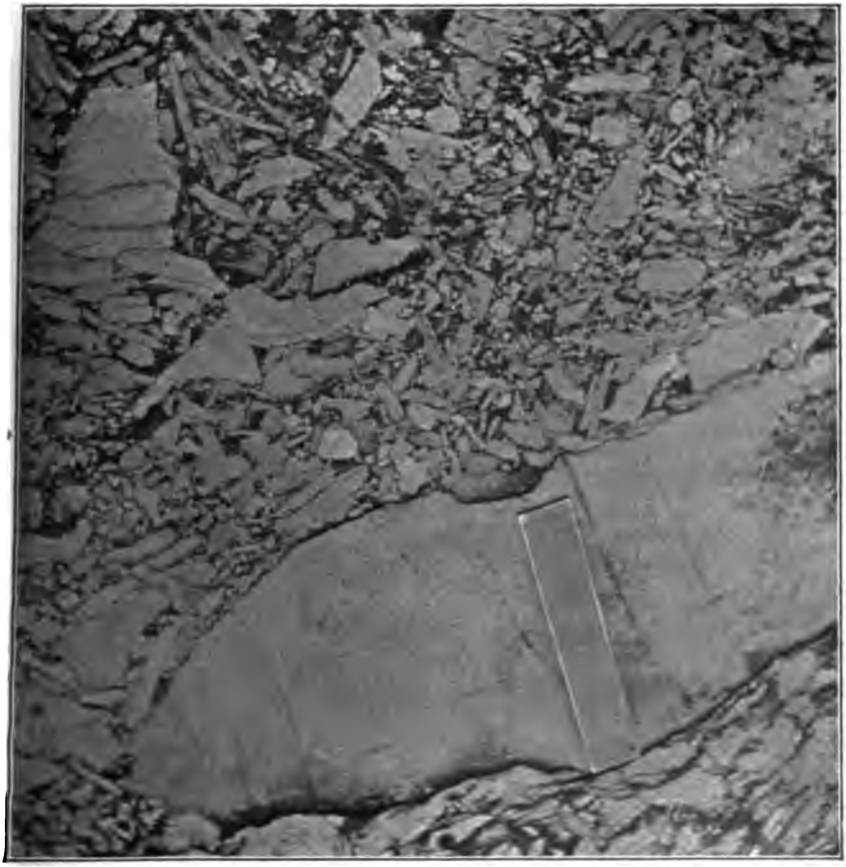

FIG. 347.—Brecciated limestone, Calciferous formation. One mile south of Highgate Falls, Vt.

reenstones embrace the diorites, dolerites, some of the gabbros and he basalts, and may even extend to the peridotites and some of the more ornblendic of the granitoid rocks. Another convenient name is *trap*. hich may be used for any dark, heavy igneous rock. The name (from *appe*, stairs) refers to the step-like arrangement which the edges of he superimposed sheets of lava often take, especially when the lava of the free-flowing, basaltic kind.

The term basalt is sometimes used to embrace any of the very fine-grained dark igneous rocks. In such cases, it covers the very fine-grained dolerites, diorites, peridotites, etc. The term granite was used originally for any coarse-grained crystalline rock, and there is a tendency to revive this early use. In general descriptions, some of our best petrographers call any coarsely crystalline rock (e.g., coarse-grained syenites, diorites, gabbros, etc.) granite. The term *granitoids* may be used with strict propriety to cover all rocks of this class.

DERIVATION OF SECONDARY ROCKS.

Rocks, though commonly made the symbol of the abiding, are subject to constant slow changes. Through these changes newer rocks have been derived from older ones, and still others in turn from these derivatives, and so on in an endless chain. All derived rocks are conveniently termed secondary, though they may be several generations removed from the primitive rocks, and even the primitive rocks, as we now understand them, may be themselves derived. The ordinary changes of rock are most active at or near the surface, and the processes of such change have already been discussed in part under the titles "Weathering" (pp. 54 and 110), "Erosion" (pp. 119–123 and 342–349), "Transportation" (pp. 115–119 and 354–355), and "Deposition" (pp. 177–204 and 355–363).

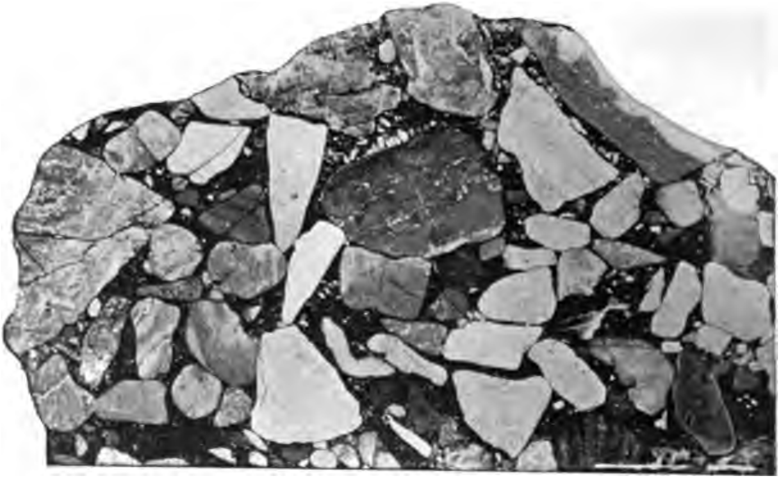

Fig. 348.—Quartzitic breccia. About one-third natural size. (Photo. by Church.)

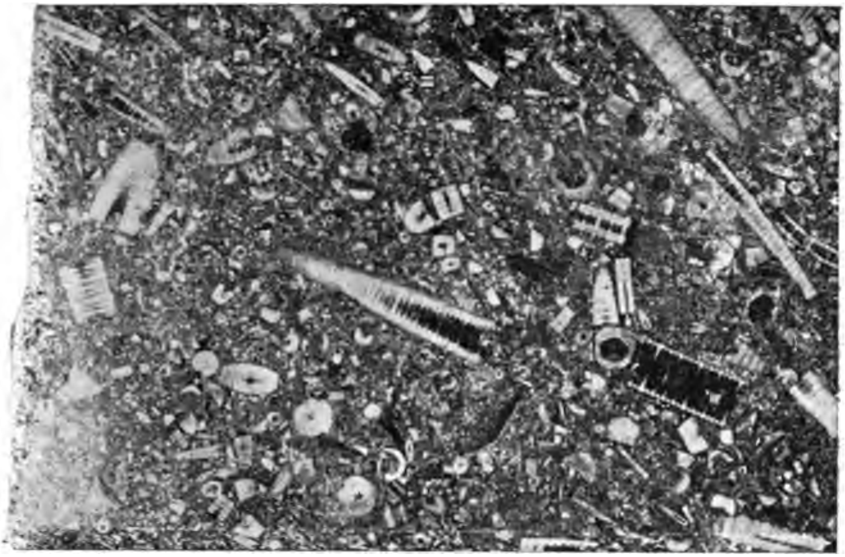

FIG. 349.—Section of limestone showing abundant fossils imbedded in a matrix made up of comminuted shell matter. About two-thirds natural size. (Photo. by Church.)

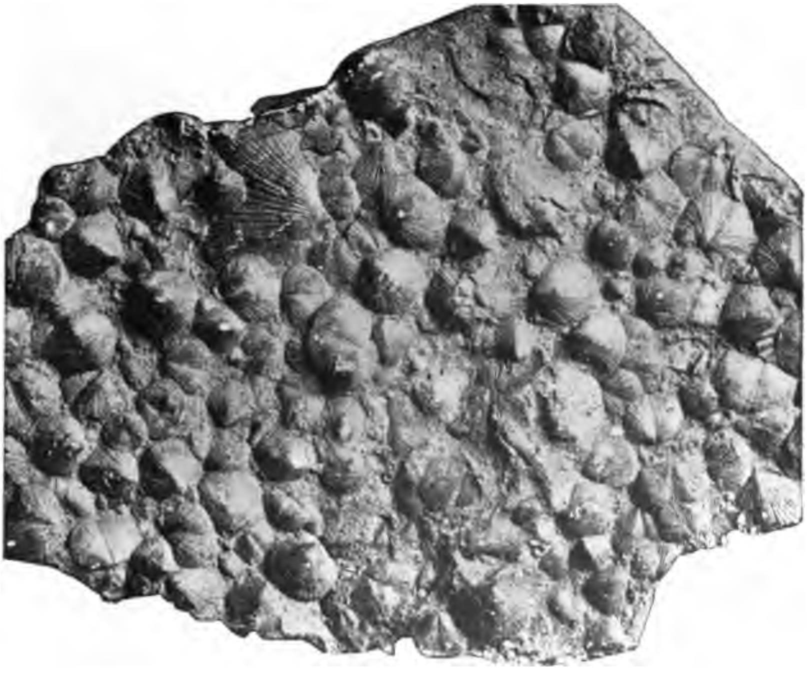

FIG. 350.—Limestone composed chiefly of shells. About three-fourths natural size. (Photo. by Church.)

Regolith.—The first great product of the surface changes is *mantle-rock* (*regolith*), which comprehends all the loose matter that springs from rock decay, wear, fracture, and other forms of disintegration. It lies in an unconsolidated sheet on the face of the land, whether as soil, sand, clay, earth, gravel, or loose rock.

Disrupted products: arkose and wacke.—In dry regions, in cold regions, on mountain heights and precipitous slopes, and under other conditions where sudden changes of temperature and frost action work efficiently, rocks are broken down into fine fragments without much chemical decomposition. Such *disaggregated* rather than decomposed matter, if derived from granitic and similar crystalline rocks, is termed *arkose*, or arkose sand, and consists of fragments of quartz, feldspar, mica, etc. Common sand consists essentially of quartz grains. If the fine fragments are derived from the darker igneous rocks, and consist mainly of grains of plagioclase feldspar, and ferromagnesian minerals, it is sometimes called *wacke*. This term is not widely used in just this sense, but there seems to be an important place for it, and it will be so employed in this work. These disaggregated sands are but special phases of the mantle-rock.

Disintegrated products.—When the surface-rock is chemically decomposed, the residual material is confined mainly to the insoluble portions, i.e., the silicious and clayey parts; while the lime, magnesia, soda, potash, and similar substances are largely dissolved and borne to the ocean. The potash is somewhat more disposed to remain with the clays than the soda, lime, or magnesia; but residues of all are usually present.

Classes of Sedimentary Rocks.

Shales, sandstones, and conglomerates.—As already shown in the discussion of the atmosphere and surface-waters, the mantle-rock is constantly being borne away and redeposited in lodgment spots on the land or in the basins of the sea, while it is constantly being renewed below. It is an evanescent but ever-renewed derivative mantle. In this process of renewal, removal, and redeposition, the mantle material is usually assorted into mud, sand, and gravel; and these several classes of material are laid down more or less separately, and usually take the stratified form, because their deposition depends on different degrees of motion of the transporting waters or wind. When these several classes of material become cemented or otherwise hardened, they give rise to *shales* (cemented muds), *sandstones* (cemented sands), and *conglomerates* (or pudding-stones, cemented gravel, Fig. 346). If the coarse material remains angular, they form *breccia* instead of conglomerates (Figs. 347 and 348). For the most part, the deposits of mud, sand, and gravel are made under the sea or in lakes and estuaries, but they are also formed on the land in lodgment basins, in low-gradient valleys, and on base-plains. The deposits of sediment on land have received less recognition than they deserve. When formed under the sea or in other life-sustaining waters, shells and other organic material are liable to be entrapped and to form a part of the rock. These organic remains, or fossils, greatly aid in interpreting the deposits in which they occur. Fossils are less liable to be preserved in sedimentary deposits formed on land. There is, therefore, some ground to suspect that great series of sandstones and shales which do not contain marine or fresh-water fossils were formed in lodgment basins on land,

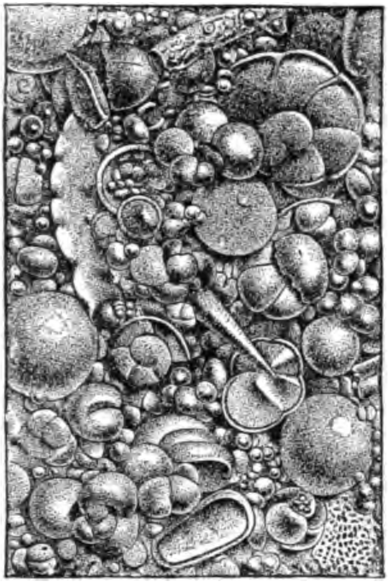

FIG. 351.

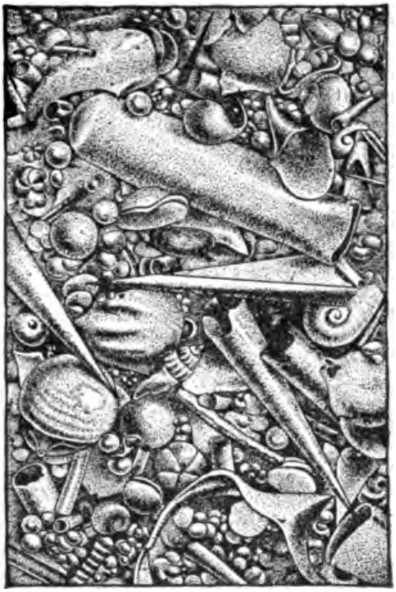

FIG. 352.

FIG. 351.—Globigerina ooze. Magnified 20 times. (Murray and Renard.)
FIG. 352.—Pteropod ooze. Magnified 4 times. (Murray and Renard.)

though the absence of fossils cannot be regarded as proof of such origin.

Limestones and dolomites.—Of the lime, magnesia, soda, and potash leached out of the surface-rocks and carried to the ocean in solution, the lime is largely extracted to form the shells, skeletons, teeth, armor, and other hard parts of sea-animals and sea-plants. These limy parts are at length left on the floor of the ocean and become more or less disintegrated and help to form beds of lime-mud and lime-sand which in time are cemented into *limestone* (Figs. 349, 350, 351, and 352). A larger proportion of the magnesia remains in solution in the sea-water, but in ways not yet well understood, the magnesia sometimes unites with the lime to form *dolomite*, a double carbonate of lime and magnesia $(Ca,Mg)CO_3$. This change is sometimes local, and sometimes affects great series of beds, more commonly the ancient ones than the modern. Sometimes the dolomization appears to have taken place long after the original limestone was formed and probably sometimes after it was lifted out of the sea, while in other cases it seems to have taken place while the sediment was accumulating, or at least before the next overlying beds were laid down. The potash in solution is to some large extent taken up by the land- and sea-plants or is retained in the clays, and through them becomes again incorporated in the sediments. The soda largely remains in solution in the sea-water.

Precipitates.—When a portion of the ocean-water is isolated in a region where evaporation from the surface of the water is greater than the rainfall on it, and the inflow from the tributary basin, the lime, magnesia, soda, potash, and other dissolved substances (solutes) are concentrated until the water becomes saturated. The solutes are then precipitated in the order in which they reach the point of saturation. This order, when taken in strict and full detail, gives a very complex series, but the leading deposits are calcium carbonate (*limestone*), calcium sulphate (*gypsum*), and sodium chloride (*halite* or *rock salt*) (see p. 375). Isolated lakes in arid regions may give rise to similar deposits. It has sometimes been thought that the ancient limestones were produced largely by precipitation from concentrated sea-water. While this is probably the case in some instances and to some degree, it has not been demonstrated that the great limestone formations were made to any large extent in this way. The more accepted view is that the limestones in the main were made from organic remains. The lime in solution in the ocean is chiefly in the form of the sulphate.

Iron Ore-beds.—In a somewhat different way iron ore-deposits are formed by the precipitation of iron oxide or iron carbonate from solutions of ferrous compounds. The ferrous compounds in solution were leached from iron-bearing rocks by percolating waters. The most familiar case is that of iron-bearing springs. On exposure to the air, the iron compounds in solution undergo change, and ferric oxide is thrown down, usually forming *limonite* ($Fe_2O_3,3H_2O$), but sometimes *hematite* (Fe_2O_3). This change is common in marshes and gives origin to "bog-ore." Similar deposits take place in certain shallow lakes, and hence are known as "lake ore." Iron ore sometimes also forms at the bottom of a peaty bed or in muddy soil. In connection with the great coal formations, beds of iron carbonate (*siderite*) occur. Organic matter seems to play a great part both in the original solution and the later deposition of these ores. From certain soils and clay-beds on which the ancient coal-producing forests grew, the iron has been almost completely removed, either by the action of the roots, or more probably by organic acids arising from their decay and from the decaying vegetation on the surface. On flowing into shallow bodies of water or into marshes, the waters containing such dissolved iron compounds usually throw down their iron content either as a carbonate (siderite), or as a hydrous ferric oxide (limonite). The siderite is formed where decaying vegetation is present to furnish abundant carbon dioxide and to partially protect the iron solution from oxidation, and the limonite where free oxidation takes place. Sand, silt, clay, or calcium carbonate often accumulates with the iron precipitate, and the result is an impure deposit which becomes an *ironstone*. Such deposits often become segregated into nodules, as will be explained later. It is thought that diatoms sometimes aid in the deposit of iron ore in shallow waters.

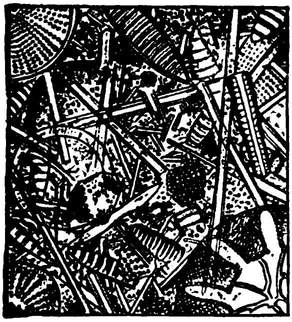

FIG. 353.—Diatom ooze. Magnified 150 times. (Murray.)

Silicious deposits.—In the decomposition of igneous rocks, a certain portion of the silica, as well as of the bases, is dissolved and carried away in solution. Certain organisms extract this from solution for their skeletons, just as others extract calcium carbonate. The accumulation of these silicious skeletons often forms silicious rocks. The *diatom*, *radiolarian*, and other oozes (Fig. 353) of the deep sea are the great examples. Sometimes layers of infusorial earth, *tripolite*, arise from the shells of diatoms and other aquatic organisms secreting silica. The waters in which such earths accumulate are rather shallow, and either fresh or salt. The most familiar examples of indurated rocks formed in this general method are the *flints* and *cherts* (impure flints) that occur in limestone and chalk, chiefly as nodules, but sometimes in distinct beds.

Organic rocks.—While most limestones, chalks, flints, cherts, and the silicious and calcareous oozes are formed through the agency of organisms, they are not themselves strictly organic. There is, however, a small but important group of rocks formed directly from organic matter. In favorable situations the woody parts of plants, falling into water, are so far preserved from decay that they accumulate in beds, and by slow changes pass into *peat*, *lignite*, *bituminous coal*, *anthracite*, and *graphite*. The first of these is composed essentially of carbohydrates and hydrocarbons much as plants are, while the last two are mainly carbon, and the intermediate members represent stages of passage from the first to the last. They are all derived from the strictly organic part of the plants, and spring essentially from the atmosphere and hydrosphere. They are only indirectly associated with the evolutions of the inorganic series.

INTERNAL ALTERATIONS OF ROCKS.

Besides the extreme alterations of rocks at the surface of the earth by which they pass into solution and into residual mantle-rock, and at length by tranportation and re-sedimentation become stratified rocks, as just described, those rocks which are not at the surface are subject to changes that give rise to several varieties of *altered rocks*. These changes are taking place constantly under ordinary conditions, though usually very slowly. Under great pressure and heat the changes are relatively rapid and intense, and lead to results not reached under other conditions. These more profound changes are termed *metamorphism*, and will be considered later. It is, however, important to recognize the great fact that the outer part of the earth, for perhaps 20,000 or 30,000 feet, is more or less fractured and permeated by water containing in solution various substances dissolved from the atmosphere, the soil, and the rocks through which it has already passed, and that this permeating and circulating water is now, and for long ages has been, working changes in the rocks, partly by dissolving matter out of them, partly by depositing matter in them, and partly by furnishing a medium through which new combinations of their constituents may take place. This outer fractured portion of the lithosphere has been called the *zone of fracture.*[1]

Oxidation and deoxidation.—At and above the surface of the underground water, where the rocks are easily reached by atmospheric waters carrying much free oxygen, and by the air itself, *oxidation* prevails. Through oxidation the ferrous oxides are changed to ferric oxides, a change which is usually manifested by a transition from a gray, green, or blue color, to buff, brown, yellow, or red. The partial progress of such oxidation is often shown in a fractured block or bowlder whose exterior shows the latter colors, while the interior shows the former. The sulphides, of which common pyrites (FeS_2) is the most familiar, are oxidized into sulphates, and then sometimes pass on into the higher oxides and other compounds. Thus copperas ($FeSO_4$) arises from pyrites (FeS_2) by direct oxidation of both Fe and S. The sulphuric acid of this compound, uniting with some base stronger than the ferrous oxide, gives rise to further oxidation and results in hematite (Fe_2O_3) and limonite ($Fe_2O_3,3H_2O$). In general, the mineral constituents of the rocks in this upper zone take on their maximum states of oxidation. This oxidation affects more or less profoundly the character of the rock as a whole. Deeper in the earth oxidation is less prevalent, and the action is sometimes reversed and deoxidation takes place. So also wherever organic matter is undergoing decomposition deoxidation is likely to occur.

Solution and deposition.—Solution preponderates in the upper part of the zone of fracture, but deposition is prevalent in its deeper parts. The calcium carbonate and silica dissolved near the surface are often deposited below as calcite and quartz. The sulphates and other sulphur compounds that are formed and dissolved near the surface are apt to be changed into sulphides lower down by deoxidation. The soluble oxides and other compounds formed near the surface are often likewise precipitated below. This is particularly true where the descending waters encounter decomposing organic matter, and where they mingle with waters that have followed other routes and have become charged with different solutes. On coming together, reaction between the constituents takes place, resulting sometimes in new solutions and sometimes in precipitation.

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[1] Van Hise. 16th Ann. U. S. Geol. Surv., Pt. I, pp. 589–94.

If these lower deposits of calcite, quartz, sulphides, etc., are made in the pores of the rock, they change its texture and composition. If they are made in fissures they constitute *veins*, and if a sufficient percentage of the vein matter consists of valuable metallic compounds, they constitute *ores*.

As the waters descend they suffer greater and greater pressure and some increase of temperature, and these changes modify their power to hold substances in solution. In general, the waters increase in solvent power, but the effect is different for different mineral substances, and hence as a rule the waters are taking up some substances and laying down others as they proceed. After penetrating to greater or less depths, the waters may come again to the surface, either because they are pushed up by the higher head of the waters behind, or because they become warmer and thus lighter, and are forced up by the heavier cold waters above, or else they pass up by diffusion through the descending waters. In any case, the deep, warm waters, usually rather highly charged with material dissolved in their previous courses, are apt to deposit some of their burden as they ascend to horizons of lower pressures and temperatures. They are particularly liable to make deposits where they commingle with other waters differently charged with solutes. Thus internal changes in the body of the rocks are, and for ages have been, taking place. In the upper part of the depositing zone, calcite is the dominant mineral deposited, while in the lower, quartz is more common; but much depends on local conditions and other influences, and no rigid rule holds good.

Hydration and dehydration.—Water sometimes unites directly with some of the constituents of a rock and produces hydrated minerals, i.e., minerals that have water as an element of their constitution, not simply water absorbed into their pores. A large class of minerals known as *zeolites*, because they swell up and undergo life-like contortions when their basic water is driven off by heat, are examples of hydrous products. A more familiar example is limonite ($Fe_2O_3,3H_2O$), of which yellow ocher is a variety, which on heating sufficiently gives off its water and becomes hematite (Fe_2O_3) or red ocher. The turning of yellow clay to red brick on burning is a familiar example of dehydration. The general tendency in the upper zones penetrated by water is toward hydration. In the lower zones, where the pressure is great, Van Hise holds that there is a tendency toward dehydration, if the rocks have been previously hydrated. This may be the case if rocks have once been near the surface and later deeply buried by the accumulation of sediments on them. If the principle holds, rocks subjected to intense lateral pressure may be dehydrated.

Carbonation and decarbonation.—The igneous rocks are largely silicates. The carbonic acid of the surface-waters and of the air acting upon them, converts them, in part, into carbonates. In this way has arisen most of the original supply of calcium and magnesium carbonates. Original carbonates formed in this way are precipitated and redissolved again and again. The carbonates in river-waters are much more largely solutions of previously solid carbonates than original carbonates formed from the silicates. The potassium and sodium of the silicates also form carbonates, but by preference they unite with the sulphur and chlorine, and hence appear more largely as sulphates and chlorides.

Carbonation is usually accompanied by oxidation and hydration. These several processes break up the complex and relatively unstable silicates into simpler and more stable silicates, carbonates, and oxides. This is illustrated by the following formulas illustrative of the changes undergone by augite and labradorite, two common rock-forming minerals.

The composition of augite may be represented by the formula

$$\left\{\begin{array}{l}\mathrm{CaO.(Mg,Fe)O.2SiO_2}\\ \mathrm{(Mg,Fe)O.(Al,Fe)_2O_3.SiO_2}.\end{array}\right.$$

Assuming Mg and Fe to be equal in amount in the first half of the above formula, and Mg and Fe to be equal in the first part of the second half, and Al and Fe to be equal in the last part of the second half, doubling the whole and allowing it to be acted on by CO_2 and H_2O, we have

$$\begin{aligned}&\mathrm{2CaO.2MgO.2FeO.Al_2O_3.Fe_2O_3.6SiO_2+6CO_2+2H_2O}\\ &\quad\mathrm{=2CaCO_3+2MgCO_3+2H_2O.Al_2O_3.2SiO_2+2FeCO_3+Fe_2O_3+4SiO_2}.\end{aligned}$$

The hydrous silicate of the last part of the equation is kaolin.

The composition of labradorite is represented by the formula

$$\left\{\begin{array}{l} CaO.Al_2O_3.2SiO_2 \\ Na_2O.Al_2O_3.6SiO_2. \end{array}\right.$$

Assuming the two molecules represented by this formula to be equally abundant, and allowing the whole to be acted on by H_2O and CO_2, we have

$$CaO.Na_2O.2Al_2O_3.8SiO_2 + 4H_2O + 2CO_2 \\ = CaCO_3 + Na_2CO_3 + 2(2H_2O.Al_2O_3.2SiO_2) + 4SiO_2.$$

When waters charged with carbonates descend into the earth they are likely to precipitate a portion of their burden, forming calcite and other crystalline carbonates, and hence these are among the most common minerals found in veins and rock cavities. Carbonates are also deposited when carbonate-charged waters come to the surface and evaporate or lose a part of their carbon dioxide.

Decarbonation also takes place, but it is, at least at the surface, a much less common process, and its conditions are less well understood. Sufficiently high heat will drive off the carbon dioxide, as in the artificial process of burning lime, but this is rarely observed in nature. Even lava intrusions do not usually reduce limestone to caustic lime at any appreciable distance from the contact. It is believed, however, that in the deeper zones, where high pressure and heat prevail, carbonates are changed into silicates, thus in a way reversing the process that prevails at the surface, and setting free again a portion of the carbon dioxide that had become locked up in the formation of the carbonates. To this action some of the carbon dioxide of deep-seated thermal springs is assigned.

The carbonation of the silicates takes place at the expense of the carbon dioxide of the atmosphere and hydrosphere, and hence in proportion as the igneous rocks are changed into carbonates, the atmosphere and hydrosphere are depleted of carbon dioxide, new supplies being neglected. As plants are dependent on carbon dioxide for their principal food, and as animals are dependent on plants for their food, directly or indirectly, the process of carbonation has a profound bearing on the life-history of the earth, and will often invite attention in the historical chapters. It is sufficient here to note that carbonation is one of the chief processes in the alteration of igneous rocks and furnishes, directly and indirectly, a larger percentage of the mineral substances dissolved in the waters that flow from the land, than any single process.

Molecular rearrangements.—Besides these and similar changes that involve additions and subtractions through the agency of percolating water, the molecules of some of the rock constituents rearrange themselves, or the elements enter into new chemical relations; thus, pyroxene may pass into hornblende by a change of the crystalline arrangement of the molecules. The change may sometimes be caught in progress, the outer part of the crystal being hornblende (which when thus formed is called *uralite*), while the heart of the crystal remains pyroxene. So aragonite may pass into calcite.

By changes of the foregoing kinds, many crystalline rocks are much altered. Some become *chloritic* from the development of the soft, green hydrated mineral, chlorite, derived from the pyroxene, amphibole, biotite, and perhaps other silicates of the original rock. Others become *talcose* from the development of talc, a very soft, unctuous, hydrous magnesian silicate developed from the magnesian minerals of the original rock. *Soapstone* or *steatite* is a rock composed essentially of such secondary material. *Serpentine* is a rock made up of a similar secondary mineral (serpentine) apparently derived from chrysolite (olivine) and other magnesian minerals. *Epidote*, a complex lime-iron-alumina silicate, often recognizable by its peculiar pistachio-green color, is derived from other silicates, and is rather common in many varieties of crystalline rocks. *Melaphyre* is a name applied rather loosely and variously to certain altered basic rocks of the basalt family. *Diabase* is essentially an altered dolerite. Nearly all the very ancient basaltic rocks show notable degrees of alteration, even though they appear to have escaped unusual dynamic conditions since their original formation, and hence their alteration seems to have resulted chiefly from the operation of unobtrusive agencies, chief among which is the circulation of water.

The Salient Features of Rock Descent.

The foregoing processes by which primitive or igneous rocks are disintegrated and their constituents converted into fragmental material may be said to constitute *the descent of rocks* in its fuller sense. Viewed chemically, the great features of the process are (1) the breaking down of the complex silicates, and (2) the gathering of the resultant simpler silicates (mainly aluminum silicates) into the silt and clay beds, (3) the assembling of a large part of the free acidic element (the quartz) into the sand and gravel beds, and (4) the concentration of a large part of the earthy basic element (the calcium, magnesium, and iron oxides) into the calcareous, magnesian, and iron deposits, while (5) a large part of the alkaline basic remainder (the sodium, and potassium oxides) is dissolved and held in the sea-water. Physically, the great features are (1) the disaggregation of the antecedent rock, and (2) the separation from one another of products which are physically unlike, that is, the coarser from the finer, and the heavier from the lighter, and (3) the aggregation of these diverse materials in more or less distinct beds. It is to be noted that while the rearrangement of the sediments is made on the basis of their physical characters, it results in chemical differentiation as well, for the products of rock decay, which are physically diverse, are often chemically diverse as well. The physical assortment and the stratification are to be looked upon as a step in the direction of a simpler grouping of the material. On the whole, the process is descensional in character.

THE REASCENSIONAL PROCESS.

Running hand in hand with this descensional process, there has always been a reascensional process by which the coherence, the crystallization, and in some measure the complex composition of the rocks are restored. This is partially due to external mechanical agencies, but chiefly to internal chemical and molecular forces.

Two general phases of this reconstructional work are recognized. The first, simplest and most universal, is that by which the incoherent materials produced by the descensional processes, i.e., the muds, sands, and clastic materials generally, are hardened into firm, coherent shales, sandstones, and limestones, and incidentally more or less changed in composition and molecular arrangement. The second is that by which more profound changes of induration and of composition are wrought, bringing the rock back to a state resembling its original crystalline character. This is known as *metamorphism.* Often, however, it is but an extension and intensification of the more common processes of the first class. Metamorphism is essentially reconstruction.

Induration under ordinary pressures and temperatures.—All kinds of loose fragmental material, whether soils, earths, clays, sands, gravels, volcanic ashes, cinders, or other forms of clastic or pyroclastic material, may become hardened into firm rock either by *pressure,* or by *cementa-* or by both. Pressure and cementation commonly act together and aid each other. The ordinary pressures arise from the weight of the overlying material, and these of course increase with depth. Extraordinary pressures arise from the shrinkage of the earth and perhaps from other sources. The fragments of the clastic material, on being pressed together for long periods, weld more or less at the points of contact. If they are irregular, angular, or elongate, they come to interlock more or less like the fragments of macadam, and this cooperates with the welding. The process is greatly aided by water-bearing solutions of lime, silica, etc. which are deposited at the points where the fragments press upon each other. It is here that the capillary spaces are most minute and deposition is most liable to take place. Sometimes a film of mineral matter is laid down over the surfaces of the fragments and serves to bind them together. This process goes on wherever the ground-waters are in a depositing condition, just as the opposite process of disintegration takes place wherever the waters are in a solvent state. At and near the surface of the land, the waters are usually in the latter condition and disintegration is in progress, as already noted, but this is not always so. At times and places, the water from within the rock-mass may come to the surface and evaporate, and in so doing leave all its dissolved material on the surface, or within the outer pores of the mass, as cementing material. The exterior thus becomes firmly bound together, "case-hardened," as it is termed. This may be seen in the drying of a lump of mud, the exterior of which often becomes quite firm. It is seen in quarry-rock, especially sandstone, which is sometimes soft and easily worked when taken wet from the earth, but which hardens as the water—the "sap" of the quarrymen—dries out and deposits its solutes in the capillary spaces of the grains of the surface. It is obvious that it is the very last of the "sap" which contains the most concentrated solutes, and that this last remnant is held in the minute capillary spaces where the grains touch each other, and hence the last stage of drying leaves the cement at the points where it is most effective. In natural exposures of sandstone, the pores of the outer shell sometimes become almost completely filled in this way with silicious deposits, and the sandstone is changed into a quartzite.

In the sea, and in the deep water underground, the common habit of the water is to deposit more than to dissolve, though it is doing more or less of both. As a rule, therefore, loose material in these situations becomes bound more or less firmly into rock, and hence what were originally loose sand beds become *sandstones;* what were soft muds become *shales* or *limestone*, according to composition; what was gravel becomes *conglomerate;* what was chipstone becomes *breccia;* what were volcanic ashes, cinders, and lapilli become *tuffs;* and what were masses of volcanic blocks and coarse fragments become *agglomerates.*

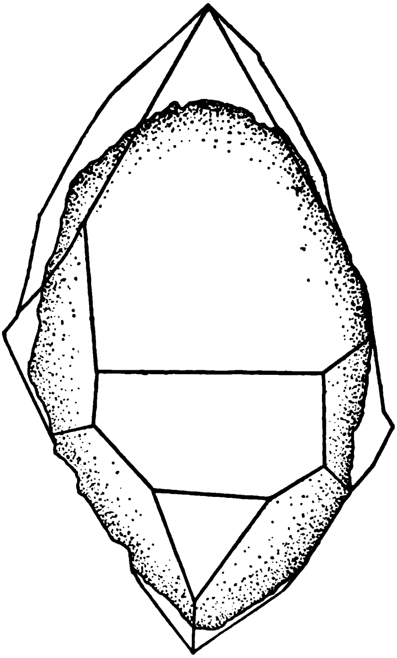


FIG. 354.

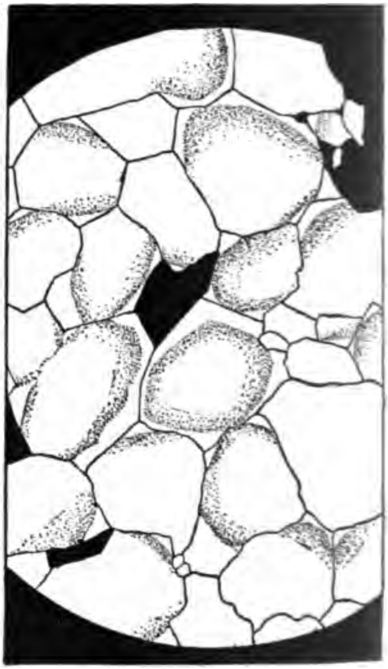


FIG. 355.

FIG. 354.—Quartz crystal enlarged by secondary growth. The shaded outline represents the outline of the sand grain; the solid lines, the outline after secondary growth. Magnified 67 diameters. (Van Hise.)

FIG. 355.—Sandstone and quartzite texture. The shaded outlines represent the surfaces of the sand grains before growth, the intervening white portions, the added quartz, and the black portions, unfilled spaces. Open spaces characterize sandstone. When the spaces are filled with quartz, the rock becomes quartzite. Magnified 35 diameters. (Van Hise.)

The cementing process works at times in specially interesting ways. In quartz sandstones, the grains are worn fragments of quartz crystals, formed originally in quartz-bearing rock. The crystalline force in these remnants controls the arrangement of the new molecules of silica de- d about them. The result is that the new deposits tend to build up the original forms of the crystals from which the sand grains were derived (Fig. 354). Sometimes a film of iron oxide has formed about the grain of sand before the addition of the new silica. This, or some difference of color, may clearly distinguish the original grain from subsequent additions. Sometimes the adjacent grains of sandstone are rebuilt in this way until the interstices are completely filled. When this has been accomplished, the sandstone becomes a quartzite (Fig. 355).

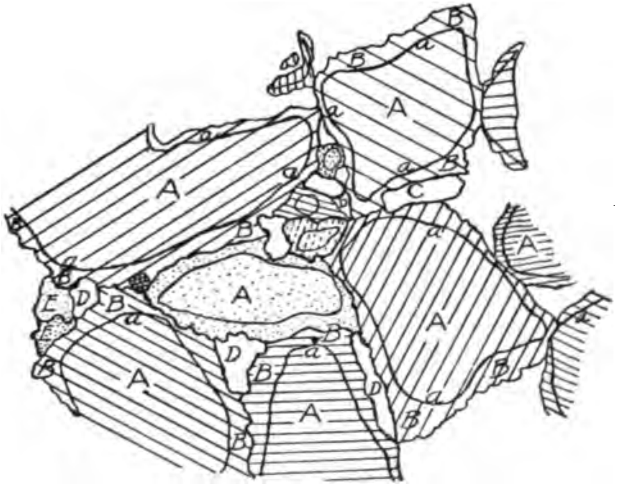


FIG. 356.—Feldspar crystals enlarged by secondary growth. Magnified 50 diameters. *AA* = original grains; *BB* = enlargements; *D* = unfilled spaces. (Van Hise.)

Most quartzites indeed appear to have been formed in this way, but mainly under special conditions that promote the deposition of silica. Grains of other minerals, such as feldspar, are subject to similar secondary enlargement (Fig. 356).

Sometimes the new material is deposited in the form of concentric shells about the particles of sediment, building them up into little spheres. Rock formed of such spherules is known as *oolite*, from the resemblance of the grains to the roe of fish (Fig. 357). Sometimes the nuclei of the concretions are grains of quartz sand, and the added concentric layers are of calcium carbonate. In this case the structure is quite obvious; but perhaps more frequently the nuclei are minute and difficult to identify, and the concentric shells make up the main mass of the grains. Certain formations, as the oolitic limestone of Indiana and elsewhere, and the Upper and Lower Oolites of England, are characterized by this structure. In most cases these accretions probably grew in depositing waters that gently rolled the grains while layers were being added. They thus do not fall under the head of cementation after the beds were formed; but concentric additions to the grains appear sometimes to have taken place after they were formed into beds.

Cavity filling.—When cavities of some size occur in rocks and the percolating waters are in a depositing state, the interiors of the cavities are sometimes lined with concentric layers of deposit. Here, instead of building *out* from a nucleus, the waters build *in* from the walls of the cavity. The *agate structure* (Fig. 358) is a case of this kind, in which the successive layers are commonly silica in the form of chalcedony and differ from each other in color and texture. Often before the cavity is entirely filled, the deposit changes from chalcedony, to crystals of quartz, which grow with their bases on the walls and their pyramidal points toward the center of the cavity. *Geodes* are examples of a similar process in which the cavity is but partially filled with crystals which have their bases set on the walls of the cavity and their points directed inwards (Fig. 359). The crystals of geodes are most commonly quartz or calcite, but they may be any other mineral that the waters are capable of

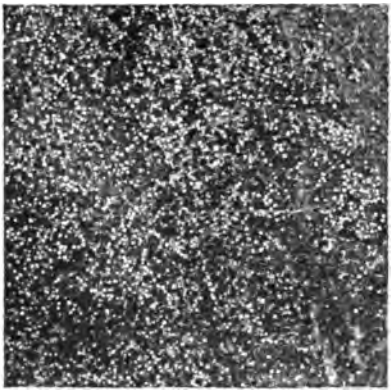


FIG. 357.—Oolitic texture. About natural size. (Photo. by Church.)

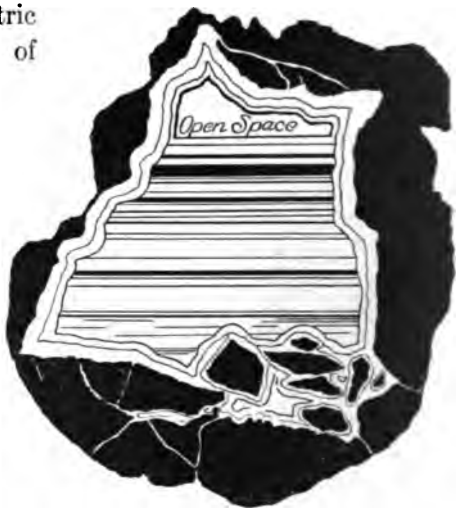


FIG. 358.—Agate structure. The cavity was first coated with mineral matter deposited from solution. The contracted cavity was then nearly filled with the same sort of material deposited in layers, apparently over the bottom, until the cavity was nearly obliterated.

depositing. Very large cavities lined in this way are known to miners as *vuggs*, and these grade on into caves lined with crystals and with *stalactite* and *stalagmite*. These are the largest expression of the solidifying process by means of internal deposition.

Fissure-filling; veins.—Cracks, crevices, and fissures filled by deposition in a similar way give rise to *veins* (Fig. 360). Here the filling grows from the walls toward the center, and hence often has a banded appearance. By this filling of cracks and crevices, the circulating water heals the breaks in the rocks. Frequently a crushed zone is thus restored to a solid state. When the fissures are deep and wide and traverse different formations, conditions are afforded for very complex deposits, and for the concentration of rare and valuable material originally dispersed through a great mass of rock. Ore deposition in such veins is usually treated as a theme by itself, but it is really but a declared expression of the work which the percolating waters are doing throughout all the rocks which they penetrate. Most of the fine crystals that grace mineralogical collections were formed in cavities and fissures by deposition from circulating mineralized waters.

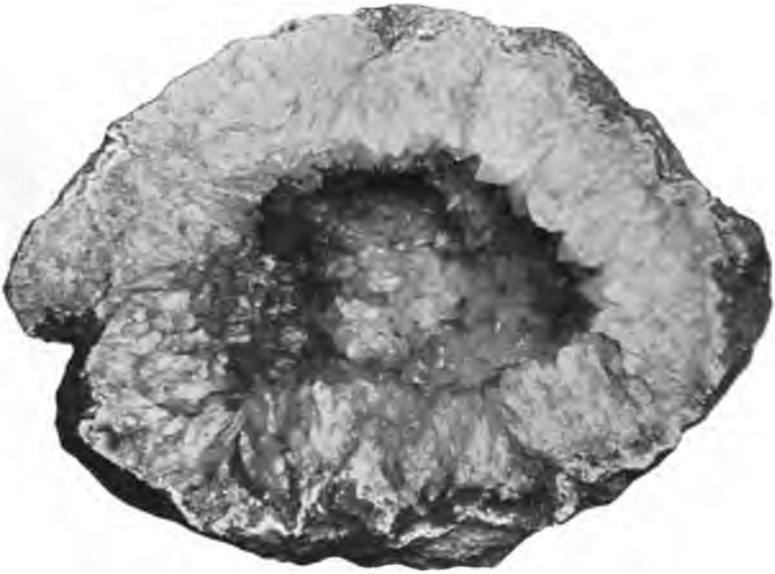


FIG. 359.—A geode. About half natural size. (Photo. by Church.)

Solution as well as deposition.—A further phase of the process needs attention. The percolating waters are constantly taking up matter as well as throwing it down, and so, while they are cementing fragments together and healing fractures, they are also removing material, and a rock may be growing porous and cavernous at the same time that its fragments are being united. Cavities may be formed at one stage and filled at another; matter may be taken up at one point and put down at another, and so an internal reconstruction is in slow progress.

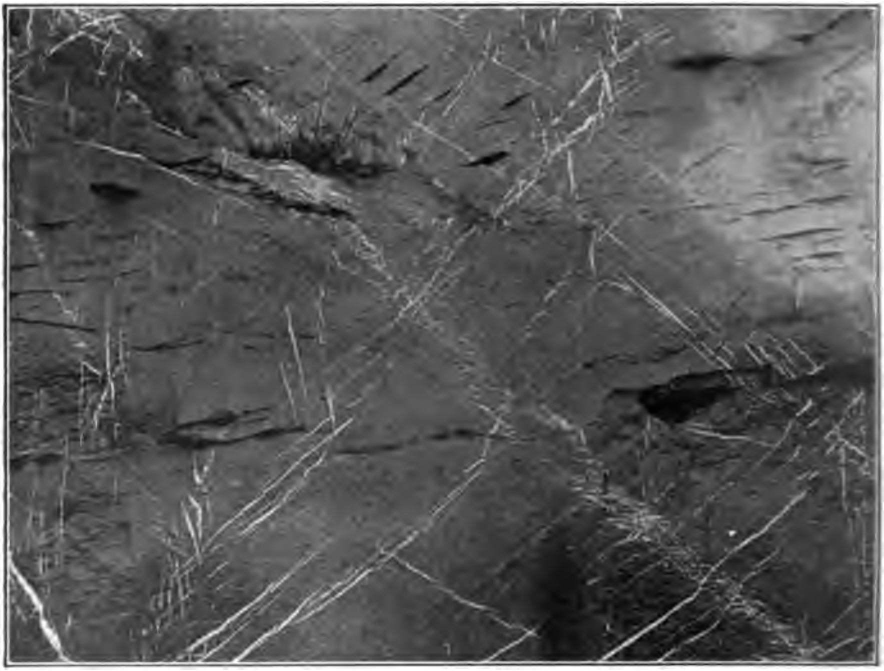


FIG. 360.—Veins of calcite in limestone. Calciferous formation near Highgate Springs, Vt. (Walcott, U. S. Geol. Surv.)

Concretions.—A notable phase of this internal reconstruction is the assembling together of like kinds of matter. For instance, silica that was probably deposited in the form of the silicious shells and spicules of plants and animals, and was disseminated through the sediments as originally formed, is aggregated into nodules of chert or flint (Fig. 361); similarly, concretions of ferrous carbonate or calcium carbonate grow in sands, silts, or muds; clusters of crystals of pyrite (FeS_2), of sphalerite (ZnS), and galenite (PbS) are formed in clayey layers, pressing the clay back as they grow; and in many other cases, kind comes to kind. Some concretions probably form during the accumulation of the beds in which they lie.

Replacements and pseudomorphs.—So also there are replacements, sometimes resulting in imitative or false forms Frequently the calcium carbonate of corals, molluscan shells, etc., is replaced by silica, and this substitution is brought about so gradually, particle by particle, that the minutest details of structure are sometimes fully preserved. This is often of great service in their study, since the limestone in which they are imbedded may often be dissolved away, while the silicified fossil is unaffected. So woody matter is sometimes replaced by silica, forming silicified wood. Similarly, the molecules of one crystal are sometimes replaced by different material, as the molecules of calcite by zinc carbonate, giving a pseudomorph of zinc carbonate after calcite.

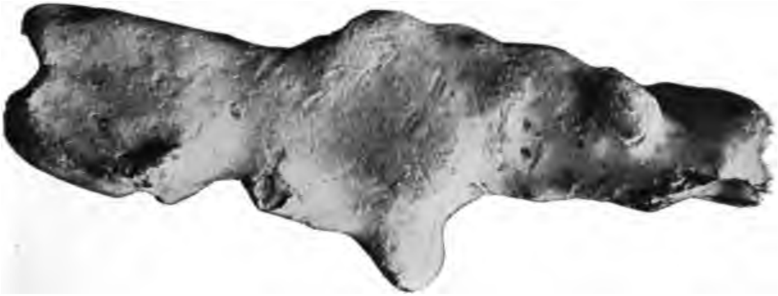


FIG. 361.—Nodule of chert. About half natural size. (Photo. by Church.)

Incipient crystallization.—A more general change is incipient crystallization. Some common limestones and dolomites are now largely made up of small crystals, though the mass was originally a calcareous mud or ooze. Incipient crystals are formed in shales and other sediments. This process, like the preceding, is a kind of incipient metamorphism or reconstruction, but it is a pervasive process, taking place under ordinary conditions of heat and pressure, and through the agency of circulating ground-waters.

By these and similar processes the fragmental deposits are solidified into firm rock and undergo internal changes which more or less reorganize the matter of which they are composed. The process is a very slow one usually. Some of the sands and muds of very early geologic ages are yet imperfectly solidified; e.g., much of the St. Peter's sandstone, a very ancient formation, is yet so incoherent as to break down into sand in being dug out, and is used for mortar sand much more than for building stone. Some of the Hudson River shales of scarcely less age are more nearly clay than hard rock. But these are examples of excessive slowness and slightness of change. In general, all but the most recent deposits show notable progress in reconstruction.

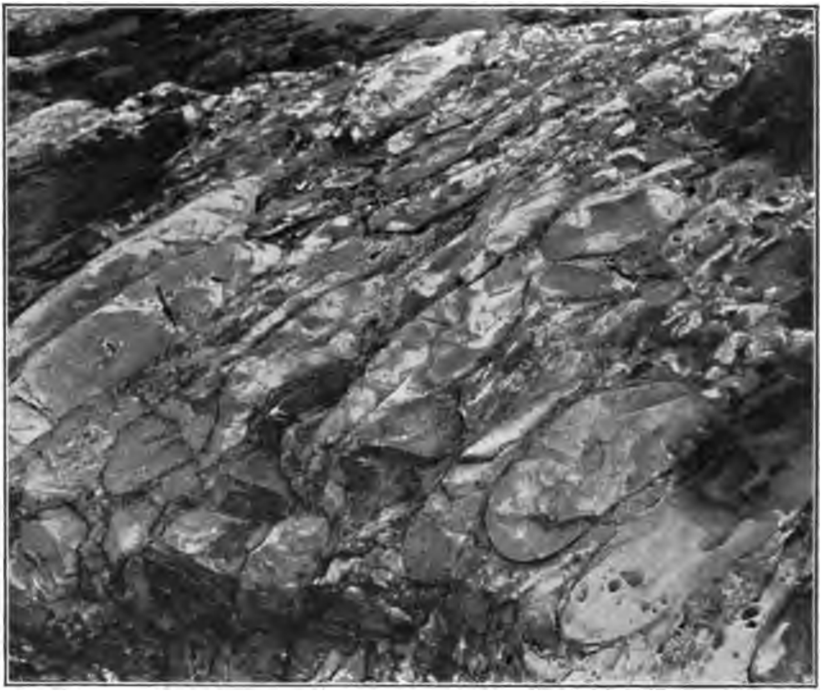


FIG. 362.—Figure showing the elongation of pebbles resulting from pressure. Carboniferous formation, Bancroft Place, Newport, R. I. (Walcott, U. S. Geol. Surv.)

Reconstruction under Exceptional Conditions.

Two special conditions greatly influence changes in rocks, viz., pressure and heat. Their action gives rise to three general cases, but these blend indefinitely: (1) exceptional pressure without great heat, (2) great heat without exceptional pressure, and (3) great heat and great pressure conjoined. Exceptional pressure may arise from the weight of overlying rocks, or from lateral thrust due to the shrinkage of the globe, and occasionally from other causes. Exceptional heat may arise from pressure, from the intrusion of hot lavas, and occasionally from other sources. In the case of intruded lavas there may or may not be exceptional pressure. Thrust usually gives heat as well as pressure, but if lateral thrust acts on rocks near the surface, they may be mashed into new forms without becoming very exceptionally heated, though some rise of temperature is inevitable.

(1) **Slaty structure.**—When rocks made up of clastic particles are compressed in a given direction and are relatively free to expand at right angles to the direction of pressure, the particles that are already elongated tend to take positions with their longer axes at right angles to the direction of pressure, and all particles, whether elongate or not, are more or less flattened in a plane transverse to the direction of pressure. This

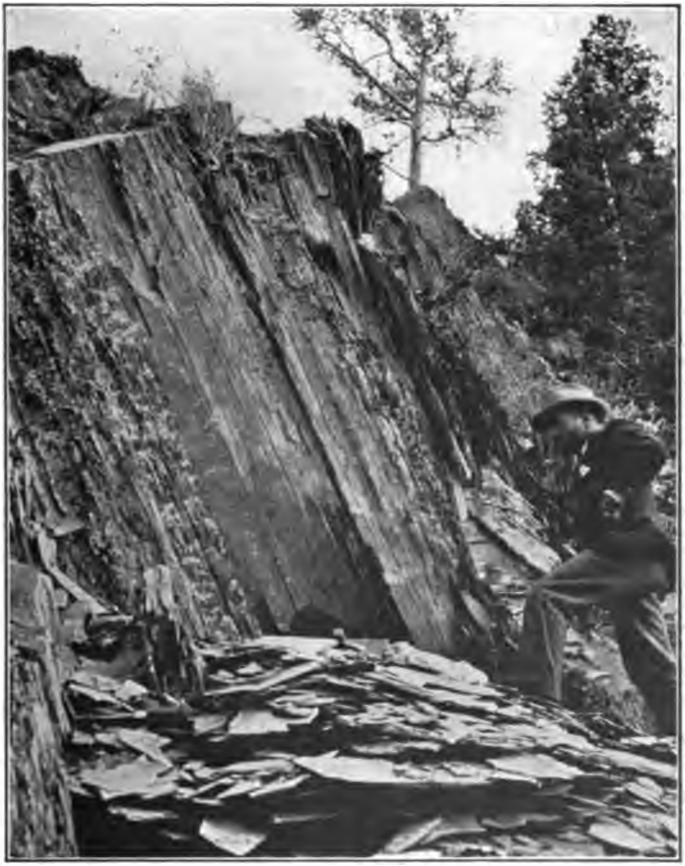


FIG. 363.—Pre-Cambrian fossiliferous slate. Deep Creek Canyon, 16 miles southeast of Townsend, Mont. (Watcott, U. S. Geol. Surv.)

may be readily seen where the particles are large (Fig. 362). As a result of the orientation and flattening of their particles, rocks so affected split more readily between the elongate and flattened particles than across them. In other words, the rocks cleave along planes normal to the direction of compression, and break with difficulty and with rough fracture across the planes of cleavage. The condition thus induced is known as slaty structure (Fig. 363), and is best illustrated by roofing-slate, which was orignally a mud, later a shale, and finally assumed the slaty condition under strong compression. Sometimes the original bedding may still be seen running across the induced cleavage planes (Fig. 364). As the original mud beds were horizontal or nearly so, and as the thrust is

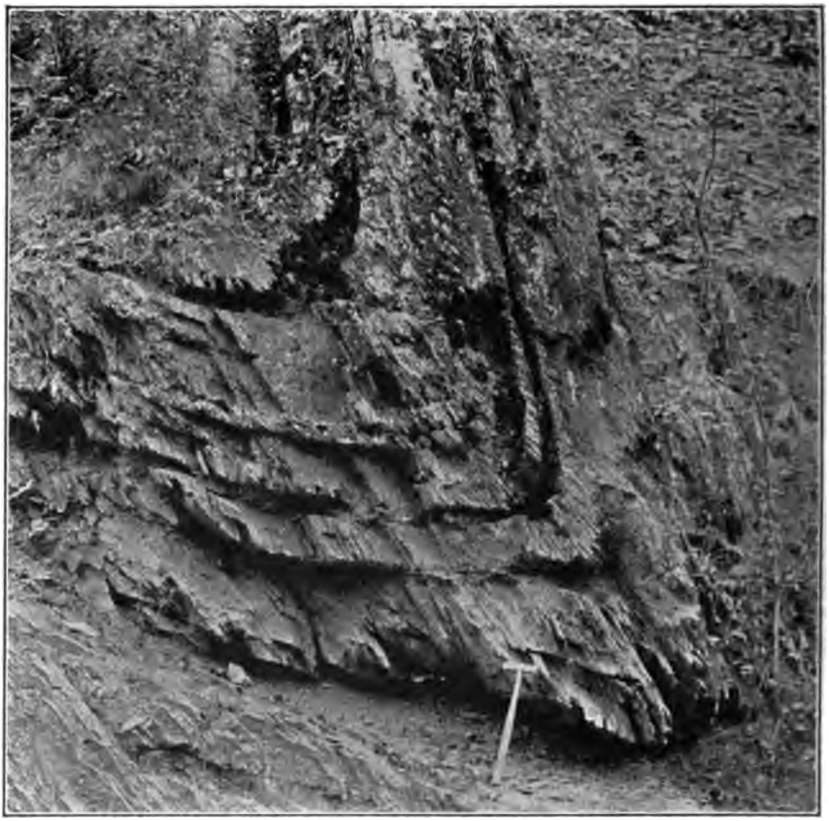


FIG. 364.—Slaty structure and its relation to bedding planes. Two miles south of Walland, Tenn. (Keith, U. S. Geol. Surv.)

usually horizontal or nearly so, the induced cleavage commonly crosses the bedding planes at a high angle (Fig. 364); but after the beds are tilted or bent, the lines of pressure take new directions relative to the bedding planes, and the angles between the original bedding and the slaty cleavages usually become smaller, and may even disappear in exceptional cases. Limestones, sandstones, and conglomerates are not so easily compressed as mudstones, and they usually take on only an imperfect cleavage normal to the direction of pressure. Often they merely show some little compacting, while the shaly strata between them are converted into slate. Obviously the direction of slaty cleavage may be used to determine the direction of the compressing force, and is thus serviceable in dynamic studies.

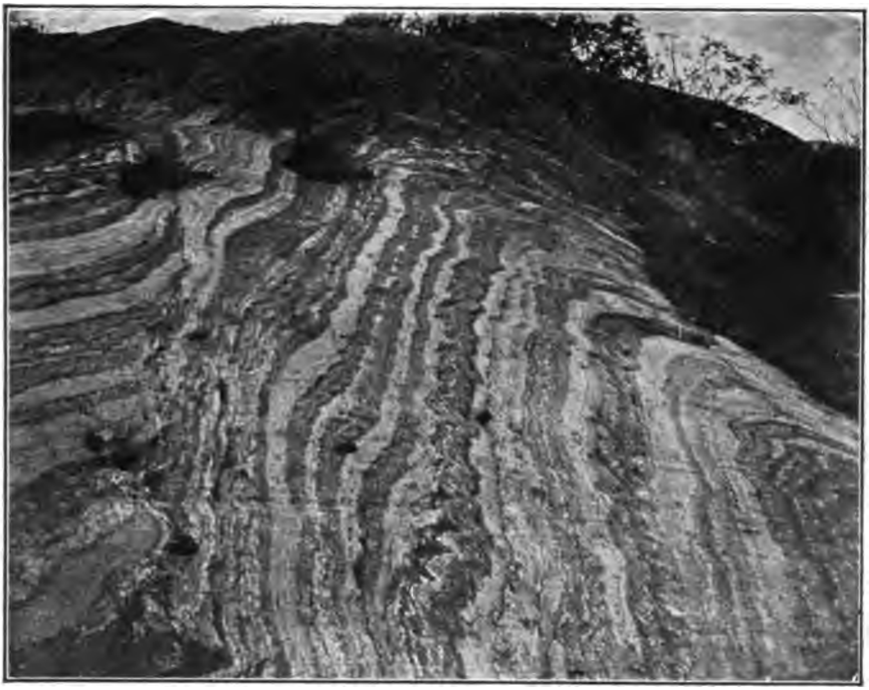


FIG. 365.—Foliated rock. (Ells, Can. Geol. Surv.)

Foliation, schistosity.—A more intense application of pressure in a given direction is capable of breaking down and deforming the most resistant rock. This must necessarily be attended with the evolution of much heat, and thermal effects are mingled with pressure effects, but the thermal effects may be neglected for the moment. The first stage of the mechanical effect of the compression may be to crush the rock more or less. It thus becomes granular or fragmental, and is really a peculiar species of clastic rock (*autoclastic*). At a further stage, the fragmented material may be pressed into layers or leaves, much as in the development of slaty cleavage, but as a result of the nature of the material, the cleavage is less perfect. This is often attended by more or less shearing of the material upon itself, and thus a rude fissility and foliation is developed. The result, including the attendant metamor-

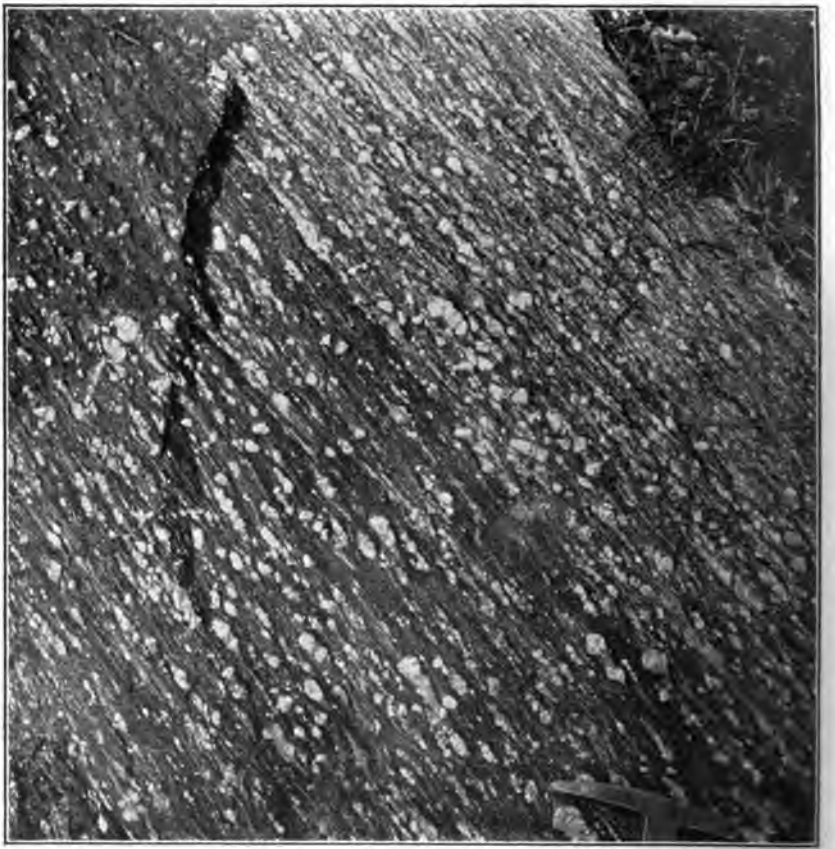
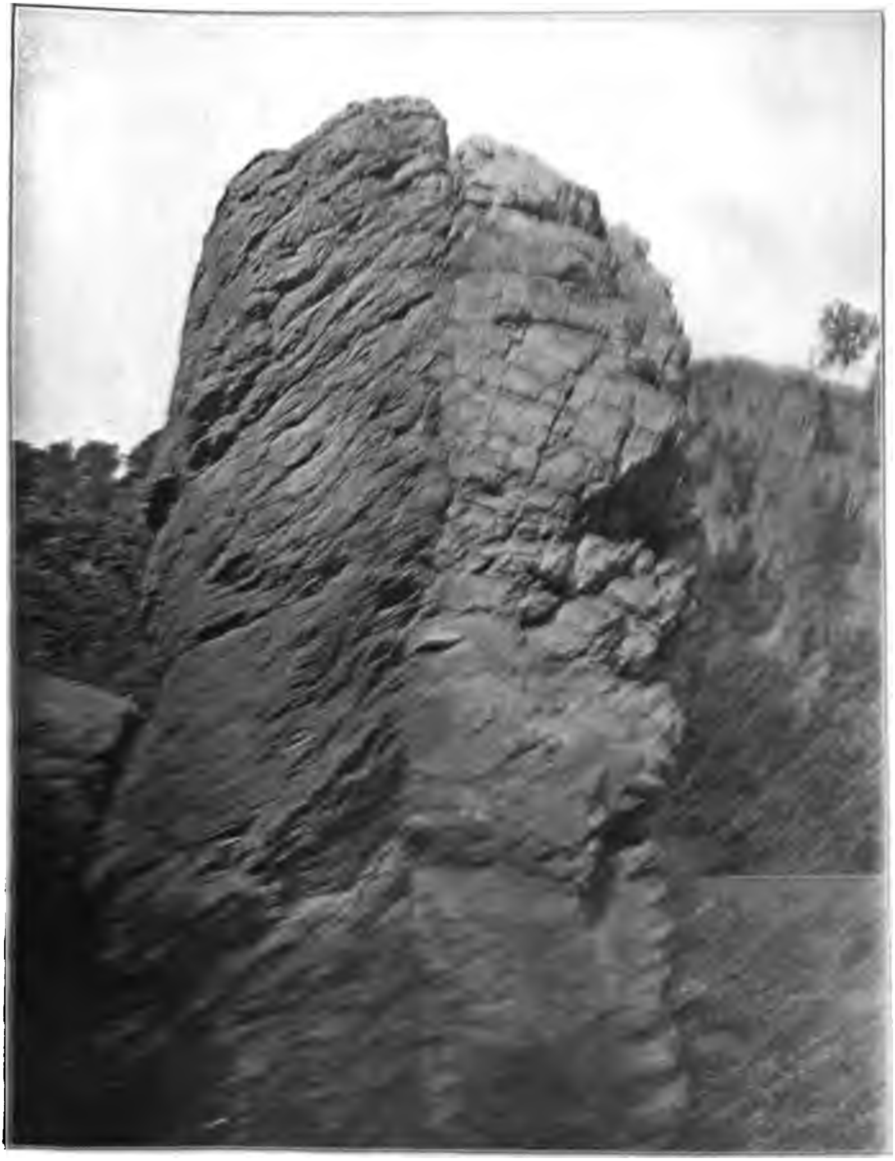


FIG. 366.—Porphyry rendered schistose by pressure. Near Green Park, Caldwell Co., N. C. (Keith, U. S. Geol. Surv.)



[illegible]

phism about to be described, is a *foliated* or *schistose structure* (Figs. 365 and 366). Even the most massive rocks may be reduced to the foliated form by this process; thus, a granite may be mashed into a *gneiss*—which is a granite in composition, but has a foliated structure—or a basalt may be converted into a *schist*, a common term for foliated crystalline rocks. Porphyritic rock rendered schistose by pressure is shown in Fig. 366. When massive rocks like granite or basalt are thus crushed down into the foliated form, the process is in a sense degradational. It is a kind of *katamorphism* or downward change. It is often difficult to differentiate the schists thus derived by degrading massive rocks, from those developed by ascensional processes from clastic formations (*anamorphism*). The action of heat is important in the evolution of schists of both classes, but the effects of heat may best be taken up where it acts measurably alone.

Metamorphism by heat.—When a mass of lava is poured out upon the surface, it bakes the mantle-rock which it overruns, in greater or less degree, depending on the mass and temperature. The nature of the effect is much the same as in the process of brick-making, a dehydration of the material, a hardening of the loose matter by the partial welding of the particles, and sometimes the partial fusion of the surface and the development of new compounds, usually glassy, but sometimes partially crystalline. In both the natural and the artificial process, the time element is short, the pressure trivial, and the water action limited. If the heat were to become sufficiently intense, the result would be fusion, i.e., a lava which would solidify into a glass. In such a case, the rock cycle would be carried back to the initial molten state and a new cycle instituted, but this does not usually take place when lava merely overflows the surface.

If lavas, instead of rising to the surface, wedge in between layers of rock and form *sills*, or interstratified sheets, the surface above as well as that below is baked, and as the excess of heat of the lava can only escape through the neighboring rock, the effects for a given mass of lava are more considerable, and as the time element and the water action (and sometimes the pressure) are usually greater than in the case of extruded lavas, the effects tend rather toward chemical and crystalline change than to simple baking. This tendency increases with increase in the mass of the lava and in its temperature. Sometimes enormous masses of very hot lava are thrust in between or among the strata that lie beneath the surface, and bring to bear upon them intense heat for a long period. So also, when a vent or fissure is the passageway for lavas that continue to come to the surface for long periods, as in the case of persistent volcanoes, the rocks which form the walls of the vent or fissure are heated for a long time, and this gives rise to metamorphism through heat, without very unusual pressure, but usually with the free aid of water. In these cases the chief effect is chemical recombination and crystallization. In the limestones and sandstones it is simple; in the shales more complex. In pure limestones and dolomites little chemical change takes place, but the molecules are rearranged into larger and more perfect crystals, and *marble* is the result. The coarseness of the crystals is, in a general way, a measure of the length of time during which the heat acts, and of its intensity, but much depends on the freedom of the attendant water circulation. Crystals an inch or two across are sometimes formed in the contact zone, where the attendant water action is important. If impurities, as silica, alumina, iron, etc., are present, various minerals, such as *tremolite* and *actinolite*, may be formed in the marble. In pure quartzose sandstones, the effect is to cause the building up of the quartz grains until the interspaces are essentially filled and the whole becomes a massive *quartzite*. Here, as in the marbles, impurities form adventitious crystals, a very common one being *hematite*, formed from the segregation of the ferric oxide of the sandstone.

In the shales, the material to be acted upon is more complex, for, while the main mass is an aluminum silicate, there is usually much free quartz, not a little potash and iron, and more or less of lime, magnesia, soda, and other ingredients, for the muds from which the shales arose contained not only the fully decomposed matter of the original crystalline rocks, but the fine matter worn from them by wind and water without decomposition. When this mixed matter is acted upon by high heat and moisture, it tends to return to its original crystalline state, so far as its changed constitution permits. The potash chiefly unites with alumina and silica, and forms potash feldspar (orthoclase chiefly) and potash mica (muscovite). The iron often unites with magnesia, alumina, and silica to form biotite or one of the ferromagnesian minerals, chiefly an amphibole. The lime usually aids in the formation of other silicates of either the feldspar or the ferromagnesian group, while the surplus silica crystallizes into quartz. There is usually a predisposition to form mica in preference to other silicates if the proper constituents are present, and the result is that *mica schists* and *gneisses*, in which mica abounds, are common products of the metamorphism of shales by contact with bodies of lava. Mica schists and micaceous gneisses are also formed in other ways, and other schists, dependent on the composition of the shales, are formed about intrusions of igneous rock. In all such cases pressure probably attends the heat and is a factor in the development of the schists. When the change induced by the heat is less considerable, the shale is baked, with incipient recrystallization, and often takes the form of *argillite*, a compact, massive sort of shale.

Beds of hydrous iron oxide (limonite) or of iron carbonate (siderite) are usually converted by heat into hematite or magnetite. Beds of peat, lignite, and bituminous coal are converted into anthracite by the driving off of the volatile hydrocarbons. If the process goes to the extreme, graphite is the result.

Metamorphism by heat and lateral pressure.—As already indicated, the more common intense pressures experienced by rocks at and near the surface are those that come from lateral thrusts arising from the shrinkage of the earth. These affect one dimension of the rock-mass, while they permit it to expand in one or both of the other dimensions. This produces a strain in all the constituent particles of the rock, and under such strain they pass more readily into solution than when free from strain, and more readily rearrange their molecules internally into positions of less strain. The crystals grow most freely along the planes of least stress, i.e., at right angles to the pressure.[1] As a consequence, where unidimensional pressure and high heat resulting from the compression unite their influence, the metamorphic changes are not only facilitated, but the rearrangement is controlled by the pressure and results in a parallel arrangement of the constituent crystals, giving a foliated or schistose character to the new rock. The changes themselves are much the same as those produced by heat and water without exceptional pressure, though some distinctions may be noted. It is to be observed, however, that two kinds of work are embraced here: the metamorphism of clastic rocks into crystalline schists, which may be regarded as an upbuilding process, anamorphism, and the mashing down of massive crystalline rocks into schists, which may be regarded as a degradational process, katamorphism. In both cases, however, there is solution and rearrangement of the molecules. The katamorphism of basalts and other basic rocks gives basic schists; that of granitic and similar rocks gives gneisses. The anamorphism of basic pyroclastic tuffs and wackes gives basic schists, while that of acid pyroclastics and most shales gives gneisses, mica schists, or similar acidic schists. It is obvious that ordinary shales cannot usually become basic schists, because in producing the original muds, the bases were generally removed; but when shales are highly calcareous and magnesian, as when they grade toward the limestones and dolomites, they may become basic schists by metamorphism, e.g., certain hornblendic schists. It is even more obvious that the limestone and sandstone formations must largely retain their distinct composition. It is thus seen that, in general, a sedimentary series anamorphosed must differ from a crystalline series katamorphosed, though both give rise to foliated or schistose rocks.

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[1] The application of these principles we owe chiefly to **Van Hise: Metamorphism** of Rocks and Rock Flowage, Bull. Geol. Soc. Am., Vol. 9, pp. 269–328.

Deep-seated metamorphism.—When the exceptional pressure arises from the weight of rocks felt at great depth, it is practically equal in all directions and the crystallization probably develops normally and is not forced into the parallel or foliated form. Rocks metamorphosed under these conditions probably tend to take the massive form rather than the schistose form, but this conclusion is theoretical rather than observational, for little or nothing is known of the history of such rocks.

Completion of the rock cycle.—The crystallizing processes of metamorphism are fundamentally similar to the processes by which rocks crystallize out of magmas, only in the first case the work is done chiefly by the aid of an aqueous solution, while in the second it is done through a mutual solution of the constituents in themselves, where water was but an incident. If the heat factor in metamorphism be sufficiently increased, aqueous solution may actually grade into magmatic solution through various degrees of softening and melting, and the cycle of changes be closed in upon itself.

VARIOUS CLASSIFICATIONS AND NOMENCLATURES.

From the foregoing sketch of the processes of rock-making it may easily be inferred that the varieties of rocks may be almost unlimited, and that they may be defined, named, and classified on many different bases; for example.

(1) If *the mode of origin* is chiefly in mind, rocks may be classed as

igneous (lavas, tuffs, etc.); *metamorphic* (schists, gneisses, anthracite, magnetite, etc.); *aqueous* (water-laid sediments, stalactites, travertine, etc.); *eolian* (dunes, loess in part); *glacial* (till, moraines); *clastic* (mantle-rock, sandstone, conglomerate, etc.); *organic* (peat, lignite, coal, etc., and indirectly, limestone, chalk, infusorial earth, etc.); and so on.

(2) If the *textural or structural characters* are in mind, rocks are designated vesicular (pumice, scoria, etc.); rhyolitic (flow-structure rocks); glassy (obsidian, tachylite); porphyritic (distinct crystals in obscure matrix); granitic (well-grained); compact, porous, earthy, arenaceous (sandy), schistose, etc.

(3) If the *chemical composition* is chiefly regarded, they may be classed as silicious, calcareous, carbonaceous, ferruginous, etc.; or, if the *chemical nature* is considered, they are grouped as acidic, basic, or neutral.

(4) If the *crystalline character* is made the basis, they are designated phanerocrystalline (distinctly crystallized), microcrystalline (minutely crystallized), cryptocrystalline (hiddenly crystallized), and amorphous (non-crystalline).

(5) If attention is fastened on *certain ingredients*, rocks are characterized as quartzose, micaceous, chloritic, talcose, pyritiferous, garnetiferous, etc.

(6) When rocks are regarded as *mineral aggregates*, if (*a*) the *aggregates are simple*, they are named from the dominant minerals, as dolomite, hornblendite, garnetite, anorthite, etc.; and if (*b*) the *aggregates are complex* they take special names, as syenite (orthoclase and hornblende), gabbro (plagioclase feldspar and pyroxene), etc.

(7) When the point of view is *structure of the mass*, they are classed as massive, stratified, shaly, laminated, slaty, foliated, schistose, etc.

(8) When *physical state* or *genesis* is considered, they are grouped as clastic, fragmental, or detrital (conglomeratic, brecciated, arenaceous, argillaceous, etc.); or pyroclastic (tufaceous, scoriaceous, agglomeratic); or massive, in a sense slightly different from that above (7).

As sometimes one of these characteristics and sometimes another is most important in a given rock, or in a given study, no one classification is satisfactory in all cases, yet each has its advantages in particular cases.

New System of Classification and Nomenclature.

The present systems of classifying and naming rocks have grown up gradually out of earlier and cruder methods, many of which were inherited from popular usage. Most of the names and definitions came into use before microscopical and other modern means of study were adopted. These systems, therefore, retain many inherited crudities and inconsistencies, and lack adaptation to present needs. They are too complex and difficult for field use and for general discussions, while not sufficiently exact and systematic for the more rigorous petrological discussions. A more adaptive and consistent practice has been earnestly sought by petrologists, and a new system of classification of igneous rocks has been offered by a group of leading American petrologists, an outline of which is here given.[1] To some extent this may be extended to the metamorphic crystalline rocks with necessary modifications and additions. The classification and nomenclature of the secondary rocks must probably always remain variable and plastic to express the various points of view which it is desirable to take. During the transition to this or some other new system, which seems inevitable, the appended alphabetical reference lists of the most common minerals and rocks, with brief definitions in accordance with current usage, will be found serviceable. The proposed system includes two parts, a *field system* and a *quantitative system*, the one applicable to rocks on casual inspection, and the other, only after detailed study.

The proposed field system.

The proposed field names are based largely on *texture* and *color*. The mineral constituents are used for subdivisions when they can be determined; otherwise they are neglected.

Classifying chiefly on the basis of texture and crystalline state, there are three groups: *Phanerites*, in which all the leading mineral constituents can be seen megascopically; *aphanites*, in which all, or at least an appreciable part, of the constituent minerals cannot be distinguished megascopically; and *glasses*, in which the material is wholly or largely vitreous.

I. The **Phanerites** may be further classified by their chief mineral constituents as follows:

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[1] Cross, Iddings, Pirsson, and Washington. Quantitative Classification of Igneous Rocks.

1. *Granites* (f.n.),[1] consisting largely of *quartz* and *feldspar* of any kind, with or without mica, hornblende, pyroxene, or other minerals. This differs from the present common use in not regarding mica as an essential constituent, and in not distinguishing between alkali feldspars and calcic feldspars, thus broadening the class.

2. *Syenites* (f.n.), consisting predominantly of *feldspar* of any kind, with subordinate amounts of hornblende, mica, or pyroxene, but with little or no quartz. This differs from the common use in giving hornblende a subordinate place, and in embracing rocks with calcic feldspars, thus broadening the class.

3. *Diorites* (f.n.), consisting predominantly of *hornblende* and subordinately of *feldspar* of any kind, with which there may be mica, pyroxene, or other minerals. This is nearly the present use except that any kind of feldspar may form the subordinate element.

4. *Gabbros* (f.n.), consisting predominantly of *pyroxene* and subordinately of *feldspar* of any kind, with or without other minerals. This nearly coincides with one of the various present uses of the term except that the range of the feldspar is increased.

5. *Dolerites*[2] (f.n.), consisting predominantly of *any ferromagnesian mineral* not distinguishable as hornblende or pyroxene, with subordinate elements of *feldspar* of any kind, and with or without other accessory minerals. A name to be used when the dominant mineral is clearly ferromagnesian, but cannot be satisfactorily identified as either hornblende or pyroxene, although it may probably be one of these. In other words, the dolerites (deceptive) embrace the whole diorite-gabbro group when too obscure for separation.

6. *Peridotites*, consisting predominantly of *olivine* and *ferromagnesian minerals*, *without* feldspar, or with very little.

7. *Pyroxenite*, consisting essentially of pyroxene without feldspar or olivine.

8. *Hornblendite*, consisting essentially of hornblende without feldspar or olivine.

II. The **Aphanites** may be *non-porphyritic* or *porphyritic*.

(*a*) Non-porphyritic aphanites when light-colored may be classed as *felsites;* when dark-colored, as *basalts*.

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[1] The initials f.n. (field names) are introduced to show that the term is used in the broad field sense proposed.

[2] Added by the authors of this work.

(*b*) The porphyritic aphanites or *porphyries*, when light-colored, are *leucophyres;* when dark-colored, are *melaphyres* (f.n.). They may be classified further, according to the kind of phenocryst imbedded in the aphanitic ground-mass, as

Quartz-porphyries, or quartzophyres;
Feldspar-porphyries, or feldspaphyres (not felsophyres);
Hornblende-porphyries, or hornblendophyres; and so on.

These may be subclassed by color, as

Quartz-leucophyres, light-colored quartz-porphyries;
Quartz-melaphyres, dark-colored quartz-porphyries;
Feldspar-leucophyres;
Feldspar-melaphyres; and so on.

III. The glasses are classified, according to color and luster, into *obsidians* or *pitchstones* when dark and lustrous; *perlites*, when a spheroidal fracture gives them a pearly appearance; and *pumice* when greatly inflated by included gases.

In general discussions, it is regarded as serviceable to use the term *granitoids* in a broad generic sense, to include all crystalline rocks of the general granitoid type, including the granites, syenites, gneisses, etc. In a similar broad way, the term *gabbroids* may be used to include the dark crystalline rocks in which the ferromagnesian minerals predominate, as the diorites, gabbros, dolerites, peridotites, etc. In this convenient and comprehensive way, two contrasted groups of igneous rocks may be designated. As the granitoids are usually acidic and the gabbroids usually basic, the grouping represents a broad fact of importance.

THE PROPOSED QUANTITATIVE SYSTEM.

The distinguishing characteristic of the more rigorous system designed to meet the needs of scientific petrology is its quantitative chemical character. All igneous rocks are classified *primarily* according to their chemical composition and only secondarily according to their mineral constituents, texture, and other characters. The rigorous application of the system requires chemical analyses of the rocks, but as these are not available in many cases, the authors of the system have devised a method of optical mineral analysis by which the nearly exact proportions of all the constituent minerals can be determined, and by knowledge of their chemical nature the results may be converted, by computation, into chemical terms. This can only be done for holocrystalline rocks whose crystals are large enough to be measured under the microscope, but aphanitic rocks may often be approximately classified by comparison with similar rocks already accurately determined. To facilitate this method of chemical analysis by measuring the minerals, the chemical composition of certain common rock-making minerals is expressed in proportional parts and tabulated, and is used somewhat as molecular weight is in ordinary chemical analysis. Certain of these are selected as *standard* minerals, the selection being such that the standard minerals embrace all the essential elements that enter into the composition of rocks. All other minerals are converted into their chemical equivalents in terms of these standard minerals by the use of the tables. All the mineral constituents being thus reduced to standard minerals, the classification is built up systematically on these standard (or standardized) minerals.

A new system of names is required, and these have been very skillfully formed by selecting significant letters from the names of the leading minerals or from words signifying their preponderance, so that short terms which carry their meaning in their forms, are secured, and this has been done so that these are usually euphonious, however strange they may seem to our preoccupied senses. For example, minerals composed chiefly of silica and *al*umina are called *salic;* those of *ferro*-*ma*gnesian minerals, *femic;* those of *al*uminous *fe*rromagnesian minerals, *alferric*, etc. When in a combination of salic and femic minerals, the salic are extremely abundant, the rock is *per*salic; if notably *do*minant, *do*salic; if the salic and femic minerals are nearly equal, *salfemic;* if the femic are *do*minant, *dofemic;* if extremely abundant, *per*femic, and so on, the system being mnemonic. This method of deriving names is applicable only to a portion of the necessary divisions. For the rest, a series of roots derived from geographic names, with a system of terminations, has been employed.

All standard minerals are divided into two groups of primary importance: one of minerals characterized by alumina, as the feldspars,—orthoclase, albite, anorthite,—leucite, nephelite, sodalite, noselite, and corundum, to which are added the closely associated minerals, quartz and zircon. This is called the *salic* group. The second group contains minerals characterized by iron and magnesia with no alumina, as hypersthene (enstatite), acmite, olivine, magnetite, hematite, and ilmenite, to which are added the closely associated minerals, titanite, perofskite, rutile, apatite, and all other rock-making minerals except those containing alumina together with iron and magnesia. The second group is called *femic.*

Aluminous ferromagnesian minerals, such as hornblende, augite, mica, etc.; are called *alferric,* and are not classed as standard minerals, because their complexity of composition makes it better to treat them as though made up of the simpler minerals of the standard list.

The composition of all igneous rocks can be expressed in terms of the relative proportions of the two groups of the standard minerals, salic and femic. By subdividing these groups successively on a mineral and chemical basis, a series of classificatory divisions of greater and greater precision has been formed. In each stage of the series, two factors only are compared, and a simple set of ratios has been selected to limit the divisions. Assuming the possiblity of a continuous range of variable mixtures of the two factors (A and B) from an extreme composed wholly of one (A), and an extreme composed wholly of the other (B), five ideal cases have been chosen as types or centerpoints about which variation in mixture may take place. These are:

$$\frac{A}{B}=\frac{1}{0},\quad \frac{A}{B}=\frac{3}{1},\quad \frac{A}{B}=\frac{1}{1},\quad \frac{A}{B}=\frac{1}{3},\quad \frac{A}{B}=\frac{0}{1}.$$

Division lines half-way between these points occur as the following ratios:

$$(1)\ \frac{A}{B}>\frac{7}{1},\quad (2)\ \frac{A}{B}<\frac{7}{1}>\frac{5}{3},\quad (3)\ \frac{A}{B}<\frac{5}{3}>\frac{3}{5},\quad (4)\ \frac{A}{B}<\frac{3}{5}>\frac{1}{7},\quad (5)\ \frac{A}{B}<\frac{1}{7}.$$

These ratios are used throughout the system. In (1) A is *extreme;* in (5) B is *extreme;* in (2) A *dominates* over B; in (4) B *dominates* over A; in (3) A and B are *equal or nearly equal.*

All igneous rocks are grouped in five (5) primary divisions called *Classes* on a basis of the proportions of the salic and femic minerals, thus:

Class I. $\frac{\text{Sal}}{\text{Fem}}>\frac{7}{1}$, extremely rich in salic minerals, called *persalane.*

II. $\frac{\text{Sal}}{\text{Fem}}<\frac{7}{1}>\frac{5}{3}$, with dominant salic minerals, called *dosalane.*

III. $\frac{\text{Sal}}{\text{Fem}}<\frac{5}{3}>\frac{3}{5}$, salic and femic minerals, equal or nearly equal, called *salfemane.*

IV. $\frac{\text{Sal}}{\text{Fem}}<\frac{3}{5}>\frac{1}{7}$, with dominant femic minerals, called *dofemane.*

V. $\frac{\text{Sal}}{\text{Fem}}<\frac{1}{7}$, extremely rich in femic minerals, called *perfemane.*

Each of these classes is divided into two *subclasses* according to the proportions of two subgroups of the preponderant group of standard minerals. Of salic minerals one subgroup includes quartz, feldspars, and the feldspathoids; the other includes corundum and zircon. Of femic minerals one subgroup includes the silicates with magnetite, ilmenite, hematite, and rutile; the other contains apatite and the remaining minerals of this group. Most known igneous rocks fall into the first subclass of each class.

The classes are further divided into *orders* according to the proportions of certain minerals in the preponderant subgroups. Thus Classes I, II, and III are each divided into nine orders on a basis of the proportions of quartz and the feldspars, and of the feldspars and the feldspathoids, quartz and feldspathoids not occurring together. The orders may be described in the same terms for each of the first three classes as follows:

Order I. $\frac{Q}{F} > \frac{7}{1}$, extremely rich in quartz, *perquaric.*

II. $\frac{Q}{F} < \frac{7}{1} > \frac{5}{3}$, quartz dominant over feldspar, *doquaric.*

III. $\frac{Q}{F} < \frac{5}{3} > \frac{3}{5}$, quartz and feldspar equal or nearly equal, *quarfelic.*

IV. $\frac{Q}{F} < \frac{3}{5} > \frac{1}{7}$, feldspar dominant over quartz, *quardofelic.*

V. $\frac{Q \text{ or } L}{F} < \frac{1}{7}$, extremely rich in feldspar, *perfelic.*

VI. $\frac{L}{F} < \frac{3}{5} > \frac{1}{7}$, feldspar dominant over feldspathoids (lenads), *lendofelic.*

VII. $\frac{L}{F} < \frac{5}{3} > \frac{3}{5}$, feldspar and lenads equal or nearly equal, *lenfelic.*

VIII. $\frac{L}{F} < \frac{7}{1} > \frac{5}{3}$, lenads dominant over feldspars, *dolenic.*

IX. $\frac{L}{F} > \frac{7}{1}$, extremely rich in lenads, *perlenic.*

In classes IV and V the preponderant minerals are femic, and in subclass 1 they are silicates, titanates, and ferrates, with hematite and rutile. These are subdivided as follows:

Silicates—pyroxenes and olivine with akermanite in one subgroup; the other minerals, magnetite, hematite, ilmenite, titanite, perofskite, rutile, in the second subgroup. This first group is called *polic*, mnemonic of pyroxene and olivine; the second group is called *mitic*, mnemonic of magnetite, ilmenite, titanite.

There are five orders in each of these classes, as follows:

Order I. $\frac{PO}{M} > \frac{7}{1}$, extremely rich in pyroxene or olivine, *perpolic.*

II. $\frac{PO}{M} < \frac{7}{1} > \frac{5}{3}$, dominant pyroxene or olivine, *dopolic.*

III. $\frac{PO}{M} < \frac{5}{3} > \frac{3}{5}$, pyroxene or olivine, equal or nearly equal to the mitic minerals, *polmitic*.

IV. $\frac{PO}{M} < \frac{3}{5} > \frac{1}{7}$, dominant mitic minerals, *domitic*.

V. $\frac{PO}{M} < \frac{1}{7}$, extremely rich in mitic minerals, *permitic*.

In the first three orders a distinction between pyroxene and olivine is recognized by sections, five in number:

Section 1. $\frac{P}{O} > \frac{7}{1}$, extremely rich in pyroxene, *perpyric*.

2. $\frac{P}{O} < \frac{7}{1} > \frac{5}{3}$, dominant pyroxene, *dopyric*.

3. $\frac{P}{O} < \frac{5}{3} > \frac{3}{5}$, pyroxene and olivine, equal or nearly equal, *pyrolic*.

4. $\frac{P}{O} < \frac{3}{5} > \frac{1}{7}$, dominant olivine, *domolic*.

5. $\frac{P}{O} < \frac{1}{7}$, extremely rich in olivine, *perolic*.

In the last two orders a distinction between the preponderant mitic minerals is recognized by suborders, five in number. The minerals containing Fe_2O_3 are compared with those containing TiO_2. The former, magnetite and hematite, are called *hemic*, mnemonic of hematite; the latter subgroup, titanite, ilmenite, perofskite, rutile, are called *tilic*, mnemonic of titanite and ilmenite. Of orders 4 and 5, there are

Suborder 1. $\frac{H}{T} > \frac{7}{1}$, hemic minerals extreme, *perhemic*.

2. $\frac{H}{T} < \frac{7}{1} > \frac{5}{3}$, dominant hemic minerals, *dohemic*.

3. $\frac{H}{T} < \frac{5}{3} > \frac{3}{5}$, hemic and tilic minerals equal or nearly equal, *tilhemic*.

4. $\frac{H}{T} < \frac{3}{5} > \frac{1}{7}$, dominant tilic minerals, *dotilic*.

5. $\frac{H}{T} < \frac{1}{7}$, tilic minerals extreme, *pertilic*.

Further subdivision, producing rangs and subrangs, is made on the character of the chemical bases in the standard minerals used in forming orders and is expressed in terms of the molecular proportions of certain oxides. For the salic minerals, forming orders in the first three classes, the bases are alkalies—K_2O and Na_2O—and lime, CaO. For the femic minerals, forming orders in the last two classes, the bases are MgO, FeO, CaO and alkalies, K_2O, Na_2O. In classes I, II, and III rangs are formed by comparing salic alkalies, $K_2O' + Na_2O'$, with salic lime, CaO′; and subrangs are formed by comparing K_2O' with Na_2O'.

Rang 1. $\frac{K_2O' + Na_2O'}{CaO'} > \frac{7}{1}$, alkalies extreme, *peralkalic*.

2. " $< \frac{7}{1} > \frac{5}{3}$, alkalies dominant, *domalkalic*.

3. " $< \frac{5}{3} > \frac{3}{5}$, alkalies and lime equal or nearly so, *alkalicalcic*.

4. " $< \frac{3}{5} > \frac{1}{7}$, lime dominant, *docalcic*.

5. " $< \frac{1}{7}$, lime extreme, *percalcic*.

Subrang 1. $\frac{K_2O'}{Na_2O'} > \frac{7}{1}$, potash extreme, *perpotassic*.

2. " $< \frac{7}{1} > \frac{5}{3}$, potash dominant, *dopotassic*.

3. " $< \frac{5}{3} > \frac{3}{5}$, potash and soda equal, *sodipotassic*.

4. " $< \frac{3}{5} > \frac{1}{7}$, soda dominant, *dosodic*.

5. " $< \frac{1}{7}$, soda extreme, *persodic*.

In classes IV and V rangs are formed by comparing femic $MgO + FeO + CaO''$ with femic alkalies $K_2O'' + Na_2O''$.

Minerals containing magnesia, iron, and lime are called *mirlic*.

Rang 1. $\frac{MgO + FeO + CaO''}{K_2O'' + Na_2O''} > \frac{7}{1}$, extremely mirlic, *permirlic*.

2. " $< \frac{7}{1} > \frac{5}{3}$, dominantly mirlic, *domirlic*.

3. " $< \frac{5}{3} > \frac{3}{5}$, equally mirlic and alkalic, *alkalimirlic*.

4. " $< \frac{3}{5} > \frac{1}{7}$, dominantly alkalic, *domalkalic*.

5. " $< \frac{1}{7}$, extremely alkalic, *peralkalic*.

Sections of rangs distinguish between $MgO + FeO$ and CaO''. Minerals with $MgO + FeO$ are called *miric*.

Section 1. $\frac{MgO + FeO}{CaO''} > \frac{7}{1}$, extremely miric, *permiric*.

2. " $< \frac{7}{1} > \frac{5}{3}$, dominantly miric, *domiric*.

3. " $< \frac{5}{3} > \frac{3}{5}$, equally miric and calcic, *calcimiric*.

4. " $< \frac{3}{5} > \frac{1}{7}$, dominantly calcic, *docalcic*.

5. " $<\frac{1}{7}$, extremely calcic, *percalcic.*

Subrangs distinguish between MgO and FeO, thus:

Subrang 1. $\frac{MgO}{FeO} > \frac{7}{1}$, extremely magnesic, *permagnesic.*

2. " $< \frac{7}{1} > \frac{5}{3}$, dominantly magnesic, *domagnesic.*

3. " $< \frac{5}{3} > \frac{3}{5}$, equally magnesic and ferrous, *magnesiferrous.*

4. " $< \frac{3}{5} > \frac{1}{7}$, dominantly ferrous, *doferrous.*

5. " $< \frac{1}{7}$, extremely ferrous, *perferrous.*

Finally a recognition of the character of the subordinate standard minerals leads to further subdivisions known as *grads* and *subgrads.* They only occur in classes II, III, and IV, because these are the only ones in which the subordinate minerals are in notable amounts. *Grads* are formed in a manner similar to that employed to produce orders. Thus grads in classes II and III correspond to orders in class IV and the reverse. Subgrads are the same in form as rangs when the difference in the treatment of salic and femic minerals is borne in mind. The names given to these divisions, which in fact recognize only the character of the magma, are derived from geographical localities and embrace many of those already in use, except that the names of orders are taken from countries or nations. Specific terminations indicate the place in the series of divisions:

ane for class, *one* for subclass.
are for order, *ore* for suborder.
ase for rang, *ose* for subrang.
ate for grad, *ote* for subgrad.

This may be illustrated as follows:

Class I. *persalane,* all rocks extremely salic.

Order 4. *britannare,* feldspar dominant over quartz, quardofelic. Many rocks of granitic composition whether crystalline or glassy.

Rang 1. *liparase,* peralkalic, rocks in which the potential feldspars are extremely alkalic, orthoclase, or albite.

Subrang 2. *Omeose,* dopotassic, rocks in which the extremely alkali feldspars are dominantly potassic, orthoclase, with subordinate albite. Examples of omeose are: granite from Omeo, Victoria, Australia, and rhyolite from Silver Cliff, Colorado.

The presence of distinctive minerals not indicated in the standard mineral composition of *norm* is expressed by qualifying the magmatic name by the name of the distinctive mineral; as, a hornblende-monzonose.

The precise texture of the rock is expressed by qualifying the magmatic name by a textural adjective; as, a grano-monzonose, a vitro-monzonose, a phyro-monzonose, etc.

REFERENCE LIST OF THE MORE COMMON MINERALS.

Actinolite—a magnesium-calcium-iron amphibole (q.v.); commonly bright green to grayish green; crystals usually slender or fibrous.

Agate—a banded or variegated chalcedony (quartz, q.v.).

Alabaster—a fine-grained variety of gypsum (q.v.), either white or delicately colored.

Albite—a soda feldspar (q.v.), an aluminum-sodium silicate; H. 5–6; cleavage perfect in two planes; luster vitreous or pearly white; occasionally bluish gray, reddish, greenish; sometimes opalescent.

Amethyst—a variety of quartz of purple or bluish-violet color, due probably to manganese.

Amphibole—the type of an important group of rock-forming minerals known as the amphibole or hornblende group; a ferromagnesian silicate, monoclinic, H. 5–6; luster vitreous to pearly; fibrous varieties often silky; black, ranging through various shades of green to light colors; embraces the magnesium-calcium varieties, tremolite and nephrite; the magnesium-calcium-iron variety actinolite; the aluminous-magnesium-iron-calcium variety *hornblende*, and others.

Analcite—analcine, one of the zeolites; a hydrous aluminum-sodium silicate; luster vitreous, colorless, white; occasionally grayish, greenish, yellowish, reddish, transparent to opaque.

Andesine—a plagioclase feldspar (q.v.); a sodium-calcium-aluminum silicate, intermediate in composition between albite and anorthite; H. 5–6; white, gray, grayish, yellowish, flesh red; luster subvitreous, inclining to pearly.

Andalusite—an aluminum silicate; luster vitreous; whitish, rose red, flesh red, variety pearly gray, reddish brown, olive-green; H. 7.5, infusible; impurities sometimes so arranged in the interior as to exhibit a colored, crossed, or tesselated appearance in cross-section (chiastolite).

Anhydrite—a calcium sulphate; H. 3–3.5; luster pearly to vitreous; white, sometimes bluish or reddish; differs from gypsum in absence of water and in its greater hardness.

Anorthite—a plagioclase feldspar (q.v.); a calcium-aluminum silicate; varies much by impurities and admixtures; H. 6–6.5; pearly or vitreous luster; white, grayish, reddish.

Anthracite—hard coal; hydrocarbon with impurities; supposed to be derived from bituminous coal by metamorphism.

Antimony—a native metal, tin-white, brittle; rather rare in native form.

Apatite—essentially calcium phosphate with chlorine or fluorine; hexagonal; H. 5; luster vitreous or subresinous; colors usually greenish to bluish, characterized by a hexagonal form.

Aragonite—a calcium carbonate; differs from calcite in *cleavage*, and in being orthorhombic; H. 3.5–4; luster vitreous or resinous; white, also gray, yellow, green, and violet.

Asphaltum—asphalt; mineral pitch, bitumen; a natural mixture of different hydrocarbons; odor bituminous; melts at 90 to 100 degrees C.; burns with a bright flame; graduates into mineral tars and through these into petroleum; probably the residue of the latter.

Augite—one of the pyroxenes (q.v.); an aluminum-calcium-magnesium-iron silicate; H. 5–6; monoclinic, crystals usually thick and stout; sometimes lamellar; also granular; black, greenish black, deep green; an important rock-forming mineral.

Beauxite—essentially hydrated alumina; occurs in concretionary grains of clay-like form, whitish to brown; valuable as a source of aluminum.

Beryl—a beryllium-aluminum silicate; hexagonal; prismatic; H. 8; luster vitreous or resinous; marl-green, pale passing into whitish; closely resembles apatite, but distinguished by superior hardness and in composition.

Barite—barites, heavy-spar, barium sulphate; orthorhombic, H. 5–3.5; luster vitreous to resinous, sometimes pearly; white, inclining to yellow, gray, blue, red, or brown; very heavy, sp. gr. 4.3–4.7.

Biotite—black mica, a potash-aluminum-magnesium-iron silicate; monoclinic; easy basal cleavage into thin laminæ; sometimes occurs as a massive aggregation of cleavable scales; H. 2.5–3; luster splendent on cleavage surface; black to dark green; cleavage surfaces smooth and shining; a very common constituent of crystalline rocks.

Bitumen—the same as asphaltum (q.v.).

Bismuth—a metal of whitish color and rather brittle nature; occurring occasionally native, usually as an ore.

Bronzite—a variety of enstatite (q.v.); grayish green to olive-green and brown with luster on cleavage surface often adamantine, pearly or bronze-like and submetallic.

Calcite—calcspar; calcium carbonate; rhombohedral, perfect rhombohedral cleavage; often taking the forms known as dogtooth spar, nail-head spar; frequently stalactitic and stalagmitic; H. 2.5–3.5; luster vitreous; white, occasionally pale shades of gray, red, green, blue, violet, yellow, brown; strong double refraction; embraces variety called Iceland spar; a very common mineral; the essential basis of limestone.

Cassiterite—tin stone; an oxide of tin; tetragonal; luster adamantine, usually splendent; brown or black, sometimes red, gray, white, or yellow; an important source of tin.

Catlinite—essentially a hardened red clay, rather a rock than a mineral; much prized by Indians for pipes.

Chalcedony—a crypto-crystalline variety of quartz having a wax-like luster, either transparent or translucent; white, grayish, pale brown to dark brown, black, sometimes delicate blue, occasionally other shades; frequently occurs as the lining or filling of cavities, taking on a botryoidal or mamillary form.

Chiastolite—andalusite (q.v.).

Chlorite—the type of an important group of secondary minerals usually characterized by a green color, softness and smoothness or unctuousness of feeling; they are usually aluminum-magnesium-iron silicates, with chemically combined water; derived from several other species, as pyroxene, amphibole, biotite, garnet, etc.; embraces a number of species, among which are clinochlore, penninite, prochlorite, and delessite.

Chromite—chromic iron; essentially an iron chromate; isometric; luster submetallic; iron black to brownish black; opaque; sometimes magnetic; resembles magnetite.

Chrysolite—olivine; essentially a magnesium-iron silicate; orthorhombic; H. 6–7; luster vitreous; green, commonly olive-green, sometimes yellow, brownish, grayish green; highly infusible; a common constituent of certain basic igneous rocks; the name olivine is more commonly used by geologists.

Chrysotile—a delicately fibrous variety of serpentine (q.v.).

Corundum—alumina; an oxide of aluminum; H. 9; rhombohedral; large crystals usually rough; luster vitreous; color blue, red, yellow, gray, and nearly white; purer forms of fine colors are *sapphires;* the red variety is *ruby*, the yellow, oriental *topaz*, the green, *emerald*, and the purple, *amethyst;* dark colors, with iron oxide, *emery.*

Delessite—a ferruginous chlorite, usually olive-green or blackish green; occurring commonly in the cavities of amygdaloids.

Diallage—a variety of pyroxene (q.v.); H. 4; characterized by thin foliæ; usually grayish green to grass-green, or deep green; luster on cleavage surface pearly, sometimes metalloid or brassy; an essential mineral in the gabbros, as sometimes defined.

Elæolite—a variety of nephelite (q.v.); occurring in large coarse crystals or massive, with greasy luster, from which the name is derived; a characteristic constituent of elæolite syenite.

Enstatite—one of the pyroxenes; essentially a magnesium silicate; orthorhombic; H. 5.5; luster a little pearly on cleavage surface; metalloidal in the bronze variety (bronzite); grayish white, yellowish white, greenish white to olive-green and brown; very infusible; a common mineral in certain basic crystalline rocks.

Epidote—a complex aluminum-calcium-iron silicate of varying composition; monoclinic; H. 6–7; luster vitreous, pearly, or resinous; color usually pistachio-green, or yellowish green to brownish green; can usually be detected by its peculiar pistachio hue, which is seldom found in other minerals; common in many crystalline rocks, usually as a secondary product.

Feldspar—a group of minerals of the first importance in rock formation, embracing orthoclase, microcline, albite, oligoclase, andesine, labradorite, anorthite, and numerous variations; aluminum silicates, with either potassium, sodium, or calcium or two or more of these; crystallizes in both the monoclinic and triclinic systems; possesses very distinct cleavage in two directions; H. 6–6.5; range in color from white through pale yellow, red, or green, and occasionally dark; triclinic feldspars frequently called plagioclase (see individual feldspars).

Fluorite—fluorspar; calcium fluoride; isometric, usually cubic; H. 4; luster vitreous, sometimes splendent; white, yellow, green, rose, crimson red, violet, sky-blue, and brown; yellow, greenish, and violet most common; occurs usually in veins or cavities in beautiful crystalline form.

Galenite—galena; lead sulphide; isometric, usually cubic; perfect cubic cleavage; luster metallic; lead-gray; a common ore of lead; occurs in veins and layers, also as linings of cavities.

Garnet—a complex silicate of varying composition, embracing aluminum, calcium, magnesium, chromium, iron, and manganese, but usually only two or three of these are present in abundance, and the varieties are characterized by the leading constituent; isometric, usually in dodecahedrons or trapezohedrons; H. 6.5–7.5; luster vitreous to resinous; commonly red or brown, sometimes yellow, white to blue, green or black; common in mica schist, gneiss, hornblende schist; also in granite, syenite, and metamorphosed limestone.

Geyserite—a concretionary deposit of silica in the opal condition; formed about geysers; white or grayish.

Glauconite—green-sand, a hydrous potassium-iron silicate usually impure, amorphous, or earthy; dull olive-green or blackish, yellowish, or grayish green; opaque, commonly occurs as grains or small aggregations.

Graphite—plumbago, black lead; a form of carbon, usually impure; rhombohedral, but rarely appearing as a crystal; more often as thin laminæ of greasy feel; yields a black adhesive powder; hence its common use for lead pencils; occurs in granite, gneiss, mica schist, crystalline limestone; sometimes results from alteration of coal by heat; occasionally occurs in basaltic rocks and meteorites.

Gypsum—a hydrous calcium sulphate; monoclinic; perfect cleavage into smooth polished plates; occurs in a variety of forms, including fibrous and granular; H. 1.5–2; luster pearly and shiny; white, sometimes gray, flesh-red, yellowish, and blue; impure varieties dark; crystallized varieties include selenite, satinspar, alabaster, etc.; easily recognized by its softness and want of effervescence with acids; occurs in beds; calcined and ground constitutes plaster of Paris.

Haüynite—a complex sodium-aluminum silicate and calcium sulphate; crystals dodecahedrons; luster vitreous or somewhat greasy; bright blue, sky-blue, or greenish blue, or green; occurs in certain igneous rocks, commonly associated with nephelite and leucite.

Hematite—ferric oxide, Fe_2O_3, iron-sesquioxide; rhombohedral, more commonly columnar, granular, botryoidal, or stalactitic; luster metallic, sometimes earthy; iron-black, dark steel-gray, red when earthy; gives red streak or powder; a leading iron ore, 70 percent. metallic iron when pure; the chief source of the red color of soils and rocks generally.

Hornblende—an amphibole; name sometimes used as a synonym for amphibole; sometimes to designate a variety under amphibole (q.v.).

Hyalite—a variety of silica in the opal condition; clear and colorless like glass, consisting of globular concretions or crusts.

Hypersthene—one of the pyroxenes; a ferromagnesian silicate; orthorhombic; H. 5–6; luster somewhat pearly on cleavage; surface often iridescent; dark brownish green, grayish, or greenish black and brown; a frequent constituent of crystalline rocks.

Iceland spar—a form of transparent calcite (q.v.).

Ilmenite—menaccanite; a titanium iron oxide; rhombohedral; resembles hematite; luster submetallic; iron-black; powder black or brownish red; occurs frequently in crystalline rocks associated with magnetite.

Iron pyrites—pyrite (q.v.).

Kaolin—kaolinite; essentially a hydrous aluminum silicate; usually in clay-like or earthy form; white or grayish white; often tinged with impurities; commonly arises from decomposition of aluminous silicates, especially the feldspars; basis of pottery and china.

Labradorite—a plagioclase feldspar; essentially an aluminum-calcium-sodium silicate; composition intermediate between that of albite and anorthite; triclinic; H. 6; luster pearly or vitreous, gray, brown, or greenish; sometimes colorless or white; frequently shows play of colors; important constituent of various crystalline rocks, especially of the basic class; usually associated with a pyroxene or amphibole.

Lepidolite—lithia mica; essentially like muscovite (q.v.) except that potash is replaced by lithia.

Leucite—essentially an aluminum-potassium silicate, allied to the feldspars; H. 5–6; luster vitreous, white, ash-gray, or smoke-gray; occurs in certain volcanic rocks, particularly lavas of Vesuvius.

Limonite—brown hematite, ocher;—a hydrous iron oxide; commonly earthy; also concretionary, stalactitic, botryoidal, and mamillary, with fibrous structure; H. 5–5.5; luster silky, sometimes submetallic, but commonly dull and earthy; brown, ocherous yellow; streak and powder yellowish brown; constitutes ocher, bog-ore, ironstone, etc.; is the chief source of the yellow color of soils and rocks; arises from the alteration of other iron ores.

Magnesite—magnesium carbonate; rhombohedral; white, yellowish, grayish white to brown; fibrous, earthy, or massive; found in altered magnesium rocks.

Magnetite—magnetic iron ore; iron oxide, Fe_3O_4; octahedral or dodecahedral; strongly magnetic; H. 5.5–6.5; abounds in igneous and metamorphic rocks.

Marcasite—white iron pyrites; iron sulphide; same composition as pyrite, which it closely resembles; H. 6–6.5; luster metallic, pale gray, bronze, or yellow; prone to decomposition; disseminated through various rocks, particularly plastic clays containing organic matter.

Martite—iron sesquioxide; originally magnetite, which by oxidation has assumed the composition of hematite.

Mica—the type of an important group of rock-forming minerals well known for their perfect cleavage into thin elastic laminæ; among the leading varieties are the common potassium mica (muscovite), the sodium mica (paragonite), the lithium mica (lepidolite), the magnesium-iron mica (biotite), the magnesium mica (phlogopite), and the iron-potash mica (lepidomelane).

Menaccanite—ilmenite; titanium iron ore (q.v.).

Microcline—a triclinic feldspar, closely resembling orthoclase in appearance and having the same composition.

Muscovite—common or potash mica; essentially an aluminum-potassium silicate; H. 2–2.5; monoclinic; remarkable for its basal cleavage; splits easily into exceedingly thin, flexible, elastic laminæ; luster vitreous, more or less pearly or silky; colorless or variously tinged brown, green, or violet; a common mineral in crystalline rocks, particularly in the granites or gneisses.

Nephelite—nepheline; essentially an aluminum-sodium silicate with potash; allied to the soda-feldspars; hexagonal; usually in thick prisms; H. 5.5–6; luster vitreous to greasy, white or yellowish, varying to greenish, bluish, and red; occurs in volcanic rocks; the variety elæolite characterizes the elæolite syenite.

Nosite—nosean; a complex sodium-aluminum silicate and sulphate, like haüynite, but with little calcium; common in phonolites.

Oligoclase—a plagioclase feldspar; essentially an aluminum-calcium-sodium silicate which may be regarded as a mixture of albite and a small amount of anorthite; triclinic; luster vitreous, pearly, or waxy; whitish grading into greenish and reddish; H. 6–7; common in crystalline rocks.

Orthoclase—a potash feldspar; essentially a potassium-aluminum silicate; varying by the replacement of the potassium by sodium and less frequently by other substitutions; monoclinic; occurring in distinct crystals and also in cryptocrystalline forms; cleavage planes perfect with pearly luster on cleavage surface; white, gray, and flesh-red, occasionally varying to greenish white and bright green; H. 6–6.5; difficultly fusible; sanidine a glassy variety; felsite a cryptocrystalline form; a very common mineral, especially in the granites and gneisses.

Olivine—chrysolite (q.v.).

Omphacite—a variety of pyroxene of grass-green color and silky to fibrous luster; allied to diallage.

Opal—silica with a varying amount of water; differs from quartz in a lack of crystallization and in lower degree of hardness; amorphous, massive; sometimes reniform, stalactitic, or tuberous; also earthy; H. 5.5–6.5; luster vitreous, inclining to resinous; white, yellow, red, brown, green, gray, blue, generally pale; colors arise from admixtures; sometimes play of colors as in precious opal.

Ozocerite—a native paraffine, mineral wax.

Petroleum—naphtha; a native mineral oil; a hydrocarbon, commonly believed to arise from organic matter, both animal and vegetable, but held by some to be due to deep-seated chemical and thermal action.

Pictotite—a variety of spinel, containing chromium.

Pisolite—a concretionary variety of calcite.

Picrolite—a variety of serpentine.

Piedmontite—a manganese epidote.

Plagioclase—a general term embracing the triclinic feldspars whose two cleavages are oblique to each other; embracing albite, oligoclase, andesine, labradorite, and anorthite (q.v.).

Plumbago—graphite (q.v.).

Psilomelane—essentially a hydrous manganese oxide occurring in massive, botryoidal, reniform, and stalactitic forms; luster submetallic; iron-black, passing into dark steel-gray; H. 5–6; the common ore of manganese.

Pseudomorph—a false form, i.e., having the form of one mineral and the composition of another; usually arises from the replacement of a mineral, particle by particle, by a solution of another substance, leaving the original form unchanged.

Pyrite—iron pyrites, fool's gold, iron sulphide; isometric; commonly in cubes; H. 6–6.5; luster metallic, splendent, or glistening; pale brass-yellow; occurs widely disseminated throughout a large class of rocks; usually harder and lighter in color than copper pyrites, and deeper in color than marcasite, which has the same composition.

Pyroxene—the type of a large and important group of rock-forming ferromagnesian minerals; varies in composition and embraces a large number of varieties; usually a magnesium-iron-calcium silicate; crystals usually thick and stout, but varying greatly; sometimes lamellar and fibrous; H. 5–6; luster vitreous inclining to resinous; green of various shades verging towards light colors, occasionally more often to browns and blacks; among the minerals belonging to the pyroxene group are augite, bronzite, diallage, diopside, enstatite, hypersthene, and others.

Quartz—crystallized silica; rhombohedral; crystals commonly six-sided prisms capped by six-sided pyramids; without cleavage; H. 7; scratches glass; usually transparent, glassy, colorless when pure, shaded by impurities to yellow, red, brown, green, blue, and black; varieties, amethyst, purple, or violet; false topaz, yellow, rose-quartz, smoky, milky, cat's eye, opalescent; aventurine, spangled with scales of mica; chalcedony is a cryptocrystalline variety; carnelian, a red chalcedony; chrysoprase, an apple-green chalcedony; prase, a leek-green variety; agate, a variegated or banded chalcedony; moss-agate, a chalcedony containing moss-like or dendritic crystallizations of iron or manganese oxide; onyx, a chalcedony in layers; sardonyx, like onyx in structure, but includes layers of sard (carnelian); jasper, an opaque-colored quartz, usually red or brown; flint, an opaque impure chalcedony; chert, an ill-defined term applied to an impure flinty rock; hornstone, a translucent, brittle, flinty rock.

Rutile—titanium oxide; tetragonal, crystals commonly in prisms; H. 6–6.5; luster metallic, adamantine; reddish brown, passing to red; sometimes yellowish, bluish, violet, and black; occurs in crystalline rocks and is a common secondary product in the form of microlites.

Sanidine—a glassy variety of orthoclase feldspar.

Satinspar—a variety of selenite or gypsum.

Selenite—a distinctly crystallized transparent form of gypsum.

Serpentine—a hydrous magnesium silicate; usually in pseudomorph forms; also fibrous, granular, cryptocrystalline, and amorphous; H. 2.5–4; luster subresinous to greasy, pearly or earthy, resinous or waxlike; feel, smooth and somewhat greasy; leek-green to blackish green and siskin green verging into brownish and other colors; apparently derived most commonly from chrysolite or olivine and also from other magnesian minerals; sometimes constitutes the bulk of rock masses.

Siderite—iron carbonate; rhombohedral; H. 3.5–4.5; luster vitreous, more or less pearly, ash-gray, yellowish or greenish, also brownish; occurs as extensive iron deposits and in crystalline rocks.

Smaragdite—a form of amphibole or hornblende (q.v.).

Spherosiderite—a globular form of siderite.

Spinel—a magnesium-aluminum oxide; crystals, octahedrons; red of various shades, passing into other colors; spinel-ruby is a variety.

Staurolite—a complex hydrous iron-magnesium-aluminum silicate; orthorhombic; disposed to cruciform shapes; occurs in schists and gneisses.

Steatite—soapstone, a variety of talc (q.v.); a hydrous magnesium silicate.

Sulphur—a well-known element occurring native in volcanic regions; also formed by the decomposition of sulphides, particularly pyrites.

Talc—a hydrous magnesium silicate; usually in foliæ; granular or fibrous forms; also compact; easy cleavage into thin flexible laminæ, but not elastic;

feel greasy; luster pearly on cleavage surface; apple-green to silvery white; H. 1–2; a secondary product from the alteration of magnesian minerals; distinguished by its soft, soapy feel, soapstone being one variety; whitish form is known as French chalk.

Titanite—calcium-titano-silicate; monoclinic; luster adamantine to resinous; brown, gray, yellow, green, and black; H. 5–5.5; occurs in various crystalline rocks.

Topaz—an aluminum silicate, with part of the oxygen replaced by fluorine; orthorhombic; H. 8; luster vitreous; colorless, straw-yellow verging to various pale shades, grayish, greenish, bluish, and reddish; distinguished by its hardness and infusibility; occurs in crystalline rock.

Tremolite—a calcium-magnesium amphibole; a common constituent of certain crystalline rocks.

Viridite—a general term used for green products of rock alteration, usually hydrous silicates of iron and magnesia; mainly chlorite.

Wad—bog manganese; a variety of psilomelane (q.v.).

Zeolite—a group of minerals derived from the alteration of various aluminous silicates.

Zircon—zirconium silicate; H. 7.5; luster adamantine; pale yellowish, grayish, yellowish green, brownish yellow, and reddish brown; infusible; occurs characteristically in square prismatic forms; found in crystalline rocks and granular limestone.

REFERENCE LIST OF THE MORE COMMON ROCKS.[1]

Adobe—a fine silty or loamy deposit formed by gentle wash from slopes and subsequent lodgment on flats; especially applied to silty accumulations in the basins and on the plains of the western dry region.

Agglomerate—an aggregate of irregular, angular, or subangular blocks of varying sizes, usually of volcanic origin, distinguished from conglomerate in which the constituents are rounded.

Alluvium—sediment deposited by streams.

Amygdaloid—a vesicular igneous rock whose cavities have become filled with minerals; the fillings are called *amygdules,* because sometimes almond-like in form.

Andesite—an aphanitic igneous rock consisting essentially of the plagioclase feldspar *andesine* (sometimes oligoclase) and pyroxene (or some related ferro-magnesian mineral); sometimes cellular, porphyritic, or even glassy; usually rich in feldspar microlites.

Anorthosite—a rock consisting mainly of the feldspar labradorite.

Aphanite—a rock whose constituents are so minute as to be indistinguishable to the naked eye; rather a condition of various rocks than of any specific rock.

Aqueous rocks—a general term applied to rocks deposited through the agency of water.

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[1] The following definitions are given, as nearly as practicable, in accordance with present common usage, which is, however, more or less varying and inconsistent.

Arenaceous rocks—either those which are mainly sand or those in which sand is a notable accessory.

Argillite—a clayey rock; usually applied to hard varieties only.

Arkose—a sand or sandstone formed of disaggregated granite or similar rock in which a notable part of the grains are feldspar or other silicate; sand when undefined, is understood to be quartzose.

Augitite—a rock mainly made up of augite.

Basalt—a dark, compact basic igneous rock consisting of a mass of minute crystals sometimes with more or less glassy base, often containing also visible crystals; composed of plagioclase and pyroxene, with olivine, magnetite, or titaniferous iron as common accessories; a basic lava in which the crystallization has taken place rapidly; usually rich in crystallites or microlites; graduates into dolerite and basic andesite.

Bituminous coal—common soft coal, intermediate between lignite and anthracite; contains much bituminous matter, i.e., hydrocarbons.

Bowlders—rounded masses of rock, particularly those that have been shaped by glaciers.

Breccia—a rock composed of angular fragments, contrasted with puddingstone or conglomerate, in which the fragments are rounded.

Buhrstone—a compact, flint-like silicious rock full of small cavities, so named from use as millstones.

Calc-sinter (calcareous tufa)—a loose cellular deposit of calcium carbonate made by springs; travertine is the better term, as tufa should be left for volcanic clastics.

Cannel coal—a very fine-grained homogeneous bituminous coal, giving off much gas and burning with a candle-like flame.

Chalk—a fine-grained soft rock composed essentially of calcium carbonate derived from minute marine organisms.

Chlorite schist—a schistose rock in which chlorite is a predominant mineral; usually greenish, whence the name.

Clastic rock—formed from the débris of broken-down rocks; the same as fragmental or detrital rock.

Clay—a term commonly applied to any soft, unctuous, adhesive deposit, but in strict use confined to material composed of aluminum silicate; many so-called clays are chiefly silicious silts or loams.

Clay ironstone—a clayey rock heavily charged with iron oxide, usually limonite; commonly in concretionary form.

Clinkstone—a name applied to phonolite because of its metallic clinking sound when struck; composed of orthoclase, with nephelite and one or more of the ferromagnesian minerals as accessories.

Chert—an impure flint, usually of light color, occurring abundantly in concretionary form as nodules in certain limestones.

Coal—a carbonaceous deposit formed from the remains of plants by partial decomposition.

Concretions—aggregates of rounded outlines formed about a nucleus; the material is various: clay, iron ore, calcite, silica, etc.

Conglomerate (pudding-stone)—a rock formed from rounded pebbles, consolidated gravel.

Coquina—a rock formed almost wholly of small and broken shells; especially applied to a shell limestone of Florida.

Dacite (quartz-andesite)—an andesite (q.v.) with quartz.

Diabase—a dolerite (q.v.) which has undergone alteration; consists essentially of plagioclase feldspar and augite, with magnetite or titaniferous iron as a common accessory; one of the greenstones.

Diatom ooze—a soft silicious deposit found on the bottom of the deep sea, made largely or partly of the shells of diatoms; similar deposits are formed from the shells of radiolaria.

Diorite—an igneous rock usually of dark-greenish color, consisting of plagioclase feldspar and hornblende; often speckled from the commingling of light feldspar and dark hornblende.

Dolerite—a fine-grained igneous rock composed of plagioclase feldspar (labradorite or anorthite) and augite (or related ferromagnesian mineral, as enstatite, olivine, or biotite), with magnetic or titaniferous iron as common accessories; crystals usually of medium size, assuming the ophitic structure; embraces many of the greenstones; graduates into basalt on the one hand and gabbro on the other.

Dolomite—a magnesian limestone.

Drift—in common American usage, a mixture of clay, sand, gravel, and bowlders formed by glacial agencies.

Eolian rocks—deposits formed by wind, embracing especially dunes and one variety of loess.

Felsite (felstone)—a light-colored aphanitic rock composed of feldspar often with quartz, in which the crystallization is very imperfect or obscure, giving a close-grained texture with conchoidal fracture and flinty aspect; certain varieties are called petrosilex and hälleflinta.

Flint—a compact dark chalcedonic or lithoid form of quartz.

Freestone—a sandstone of uniform grain without special tendency to split in any direction.

Fulgurites—glassy tubes, produced through fusion by lightning in penetrating sand, earth, or rock.

Gabbro (euphotide)—a crystalline rock composed of the plagioclase feldspar, labradorite (or anorthite), and diallage (or a related ferromagnesian mineral), with magnetite or titaniferous iron as a common accessory.

Gangue—a term applied to the crystalline material in which ores are imbedded.

Gannister—essentially a quartz silt or pulverized quartz used for lining iron furnaces.

Garnetite—a rock composed largely of garnets.

Geest—residual earth or clay left by the decomposition of rocks, especially limestones.

Geyserite—the silicious sinter deposited about hot springs.

Globulites—minute spherical bodies embraced in volcanic glass.

Gneiss—a foliated granite, consisting typically of quartz, feldspar, and mica; the feldspar typically orthoclase.

Granite—a granular crystalline aggregate of quartz, feldspar, and mica; the feldspar typically orthoclase; popularly and properly used for any distinctly granular crystalline rock.

Granitell—a name used to designate a quartz-feldspar rock.

Granitite—a biotite granite with quartz.

Granulite—a fine-grained granite with little or no mica.

Greensand—a sand or sandstone containing a notable percentage of grains of glauconite.

Greenstone—a comprehensive term used to designate igneous and metamorphic crystalline rocks of greenish hue and of intricate and often minute crystallization; they are mostly dolerites, diabases, and diorites; a convenient term for field use where the constituents cannot be determined, and for general use when the variety is unimportant.

Greisen—an aggregate of quartz and mica, i.e., a granite without feldspar.

Greywacke—a sand rock in which the grains are basic silicates instead of quartz.

Hälleflinta—a compact flint-like felsitic rock.

Hornblendite—a rock essentially composed of hornblende.

Hornstone—a very compact, silicious rock of horn-like texture, allied to flint; term also applied to flinty forms of felsite.

Hypogene rocks—those formed deep within the earth under the influence of heat and pressure.

Ironstone—a rock composed largely of iron, usually applied to clayey rocks having a large iron content.

Infusorial earth (tripolite)—an earthy or silt deposit consisting chiefly of the silicious shells of diatoms.

Itacolumite—a flexible sandstone whose pliability is due to an open arrangement of sand grains which are held together by scales of mica.

Jasper—a reddish variety of chalcedonic quartz.

Keratophyre—a felsite with a large percentage of soda.

Kersantite—a mica dolerite consisting chiefly of plagioclase, augite, and biotite.

Lapilli—small fragments of lava ejected from volcanoes; volcanic cinders.

Laterite—a red, porous, ferruginous residual earth of India and other tropical countries.

Lava—a molten rock, especially applied to flows upon the surface, whether from vents or from fissures; also applied to the solidified product.

Lignite (brown coal)—a soft, brown, impure coal.

Limburgite—a compact basic igneous rock of the basaltic class, composed essentially of augite and olivine, with magnetite iron and apatite as common accessories.

Limestone—a rock composed primarily of calcium carbonate, though magnesium sometimes replaces a part of the calcium. (See dolomite and marble.)

Liparite (rhyolite)—an acidic igneous rock of aphanitic or glassy texture, characterized by flowage lines and various microscopic crystals; rhyolite is the more common American name.

Loess—a very fine porous silicious silt containing some calcareous material which often collects in nodules (*Löss Kindchen*) or in vertical tubules; characterized by a peculiar competency [illegible]

[illegible] by others to be [illegible]

[illegible]

Peastone (pisolite)—a very coarse variety of oolite.

Peat—the dark brown or black residuum arising from the partial decomposition of mosses and vegetable tissue in marshes and wet places.

Pegmatite—a term of ill-defined usage applied to rocks whose grain varies from coarser to finer, and often takes on peculiar aspects due to the simultaneous crystallization and mutual intergrowths of the crystals; graphic granite is a distinct type of pegmatite in which quartz and orthoclase crystals grew together along parallel axes so that cross-sections give figures resembling certain Semitic letters (Fig. 345).

Peridotite—a very basic igneous rock composed chiefly of olivine with augite or related ferromagnesian minerals, with magnetite and chromite as accessories.

Pelites—a general term embracing clay rocks.

Perlite (pearlstone)—a form of glassy lava made up in part of small spheroids formed of concentric layers which have a lustrous appearance like pearls.

Petrosilex—an old name for felsite or hälleflinta.

Phonolite (nephelite-trachyte, clinkstone)—a compact resonant igneous rock formed of sanidine and nephelite with accessories.

Phyllite (argillite)—a variety of indurated, partly metamorphosed, clay silt in which finely disseminated micaceous scales are abundant and lustrous; intermediate between typical clay slate and mica-schist.

Pitchstone—a dark vitreous, acid, igneous rock of less perfect glassy texture than obsidian and more resinous and pitch-like.

Plutonic rocks—igneous rocks formed deep within the earth under the influence of high heat and pressure; hypogene rocks; distinguished from eruptive rocks formed at the surface.

Porphyrite—a term sometimes used for an altered form of andesite, usually porphyritic in structure.

Porphyry—a rock consisting of distinct crystals embedded in an aphanitic ground-mass.

Propylite—an altered form of andesite and similar igneous rocks.

Protogine—a hydrated micaceous or chloritic variety of granite or gneiss.

Pumice—a glassy form of lava rendered very vesicular through inflation by steam.

Pyroclastic rocks—fragmental or clastic rocks produced through igneous agencies, embracing volcanic ashes, tuffs, agglomerates, etc.

Pyroxenite—an igneous rock consisting essentially of pyroxene.

Quartzite—a rock consisting essentially of quartz, usually formed from quartzose sandstone by cementation or metamorphic action.

Regolith—a name recently suggested by Merrill to embrace the earthy mantle that covers indurated rocks, chiefly residuary earths; mantle-rock.

Rhyolite—an aphanitic or glassy igneous rock showing flowage lines, usually applied only to the acidic varieties.

Sandstone—indurated sand usually composed of grains of quartz, but not necessarily so; sometimes formed of calcareous grains or of grains of the various silicates.

Schist—a crystalline rock having a foliated or parallel structure, splitting easily into slabs or flakes, less uniform than slate; they are mainly composed of the silicate minerals.

Scoriæ—light, cellular fragments of volcanic rock, coarser than pumice; cinders.

Septaria—concretions the interior of which have parted, and the gaping cracks become filled with calcite or other mineral deposited from solution (Figs. 375–77).

Serpentine—a rock consisting largely of serpentine; derived in most cases by alteration from magnesian silicate rocks.

Shale—a more or less laminated rock, consisting of indurated muds, silts, or clays.

Slate—an argillaceous rock which is finely laminated and fissile, either due to very uniform sedimentation or (more properly) to compression at right angles to the cleavage planes; e.g., common roofing-slate (Fig. 362).

Soapstone (steatite)—a soft unctuous rock, composed mainly of talc.

Stalactites—pendant icicle-like forms of calcium carbonate deposited from dripping water.

Stalagmite—the complement of stalactites formed by calcareous waters dripping upon the floors of caverns.

Steatite—see soapstone.

Syenite—a granitoid rock composed of orthoclase and hornblende, or other ferromagnesian mineral; the name was formerly applied to a granitoid aggregate of quartz, feldspar, and hornblende.

Tachylite (hyalomelane, basaltic glass)—a black glass of basaltic nature corresponding to the acidic glasses, obsidian and pitchstone.

Till (bowlder clay)—a stony or bowldery clay or rock rubbish formed by glaciers.

Trachyte—a name formerly applied to a rock possessing a peculiar roughness due to its cellular structure; but at present mainly confined to a compact, usually porphyritic igneous rock, consisting mainly of sanidine associated with varying amounts of triclinic feldspar, augite, hornblende, and biotite.

Trap—a general term for igneous rocks of the darker basaltic types.

Travertine—a limestone deposited from calcareous waters, chiefly springs; usually soft and cellular, and hence also called calcareous tufa, calc sinter.

Tuff (tufa)—a term including certain porous granular or cellular rocks of diverse origins; the volcanic tuffs embrace the finer kinds of pyroclastic detritus, as ashes, cinders, etc.; the calcareous tufa embrace the granular and cellular deposits of springs; the better usage limits the term to volcanic clastics.

Waterlime—an impure argillaceous limestone possessing hydraulic properties.

Wacke—a dark earthy or granular deposit formed from basic tuffs or from the disaggregation of basaltic and similar rocks; a term which may well come into more general use to distinguish the silicate sands that arise from the disaggregation, but only partial decomposition, of basic rocks, as arkose does, the like products of the acidic or granitoid rocks, and as sandstone does, the granular products of complete chemical decomposition.

ORE-DEPOSITS.[1]

Ore-deposits are but a special phase of the rock-forming processes already discussed. They have peculiar interest because of their industrial value. An ore is simply a rock that contains a metal that can be profitably extracted, though for convenience the term is used more broadly to include unworkable lean ores and ore material. The metal need not preponderate or form any fixed percentage of the whole, for the criterion is solely economic and not petrologic. A gold ore rarely contains more than a very small fraction of one percent. of the precious metal, while high-grade iron ore yields sixty-odd percent. of the metal. In iron ore, the metallic oxide or carbonate makes up nearly the whole rock; in gold ore, the metal is the merest incidental constituent, from the petrologic point of view.

Concentration.—The essential fact in the formation of ores is the unusual concentration of the metal. There are vast quantities of all the metals disseminated through the rock substance of the earth and even throughout the hydrosphere, but they do not constitute ores because they have no economic value. They become ores when concentrated in accessible places to a workable richness. The degree of concentration required is measured by the value of the metal. The essential elements for consideration are, therefore, (1) the original distribution of the metallic materials through the rocks, (2) their solution by circulating waters (or, rarely, by other means), (3) their transportation in solution to the place of deposit, (4) their precipitation in concentrated form, and (5) perhaps their further concentration and purification by subsequent processes.

Exceptional and doubtful cases.—There are a few cases where ore-deposits are made by volcanic fumes or vapors, but these may be neglected here. Formerly, ores were often attributed to vapors supposed to arise from the hot interior, but this mode of origin seems incompatible with physical conditions. Ores have been attributed to water originally contained as steam in lavas, and to waters escaping from the interior of the earth, these waters being supposed to be especially mineralized. Direct evidence on this point is obviously beyond reach. Segregation in the molten state is recognized as a source of ores, but its function is probably confined chiefly to partial enrichment as stated below. There are other occasional methods, but the chief process of concentration, immeasurably surpassing all others, consists in the leaching out of ore materials disseminated through the country rock and their redeposition in segregated forms, as an incident of the recognized system of water circulation.

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[1] A comprehensive discussion of the "Genesis of Ore Deposits" may be found in Vols. XXIII and XXIV of the Trans. of the Am. Inst. of Min. Eng. (also printed with additions in book form by the Institute, 1902), in which Posepny, Emmons, Van Hise, LeConte, Blake, Becker, Ricard, Raymond, Lindgren, Weed, Vogt, Winslow, Winchell (H. V.), Church, Cazin, Adams, Keyes, Bain, Collins, Beck, and DeLaunay participated. Various phases of the leading modern views are set forth

Original distribution.—The original distribution of ore material through the primitive rocks is beyond the ken of present science, for even the nature of the true primitive rocks is unknown. For present purposes it is sufficient to regard all rocks concerned in ore-deposition as either igneous or sedimentary, and to inquire, as a first step, how far ordinary igneous and sedimentary processes contribute to the segregation of ore material, leaving for a second stage of inquiry the subsequent processes of concentration.

Magmatic segregation.—In a few instances workable masses of ore seem to have arisen from lavas by direct segregation in the molten state, without the aid of subsequent concentration by water action, on which most ores are dependent. It is not improbable that the segregation of metallic iron and nickel, and perhaps other metals, in the deeper parts of the earth may be a prevalent process, giving rise to masses like the native iron found in basalt in Greenland. This iron closely resembles the nickel-irons of meteorites, which may be illustrations of similar action in small planetary bodies that have been disrupted. Metallic masses so segregated presumably gravitate toward the planetary center and hence, whatever their inherent interest, have little relation to a subject whose basal criterion is economic. It is not at all improbable, however, that in the magmatic differentiation of the lavas that come to the surface, there is some metallic segregation that may make the enriched parts effective ground for the concentrating processes of water circulation, and so determine the location of ore-deposits. Igneous rocks are not equally the seats of ore-deposits, even when the circulatory conditions seem to be equally favorable. These conditions may not really be equally favorable, but there is good ground to believe that some igneous masses constitute a richer field for concentration than others. No definite rule, however, for distinguishing rich varieties of rock from lean ones has been determined. The basic igneous rocks are, on the whole, perhaps somewhat richer in ores than the acidic class, but there is no established law. Many acidic rocks bear more and richer ores than many basic ones. The view here entertained is that both classes are subject to regional enrichment through conditions connected with their origin, as yet little known.

Marine segregation and dispersion.—In the formation of the sedimentary rocks from the primitive and igneous rocks there was notable metallic concentration in some cases, and even more notable depletion in others. The ground-waters of the land, after their subterranean circuits, carried into the water-basins various metallic substances in solution. These were either precipitated early in the marine or lacustrine drift of the waters, or became diffused throughout the oceanic body. In the main they appear to have been widely diffused, and either to have remained long in solution, or to have been very sparsely deposited through the marine or lacustrine sediments. As a rule, these sediments seem to contain less of valuable ore material than igneous rocks, and this is rational, for, as we shall see, the ground-water circulation of the land tends to concentrate and hold back a part of the metallic content of the land rocks so that only a residue reaches the sea. But there are important exceptions to this general rule of sedimentary leanness.

The iron-ore beds of Clinton age ranging from New York to Alabama, and appearing also in Wisconsin and Nova Scotia, form a stratum in the midst of the ordinary sediments, and contain marine fossils. The great ore beds of Lake Superior were originally of similar type, and so are most other important iron deposits. It cannot be said, in most cases, that these iron deposits are marine as distinguished from lacustrine or lodgment deposits, but they are at least sedimentary. The ferruginous material was originally disseminated widely through antecedent land rocks, but was concentrated in the course of the sedimentary processes.

Limestone appears to have been sometimes enriched locally in lead and zinc, and more rarely in copper, in the course of its sedimentation. The lead and zinc regions of the Mississippi basin have been regarded as dependent on such regional enrichment as a primary condition. This localized enrichment has been attributed to solutions brought into the sea from neighboring metal-bearing lands and precipitated by organic action in the sea-water,[1] this organic action being more effective in some areas than in others because of the unequal distribution of life and the concentration of its decaying products. It is assumed that such precipitates were at first too diffuse to be of value, and further concentration was required to bring them together into workable deposits; but the further processes appear to have been effective only where the preliminary enrichment had taken place. At any rate, the workable deposits are singularly localized, while the concentrative processes are very general.

Metallic material is sometimes partially concentrated in sandstones and shales in the process of sedimentation, though more rarely. The copper-bearing shale (Kupferschiefer) of the Zechstein group in Germany, so extensively worked along the flanks of the Harz Mountains, is a striking example.

It is in every way reasonable to suppose that land-waters, on reaching the margins of the water-basins, must occasionally find conditions favorable for the precipitation of their metallic contents, and that the ratio of these precipitates to other material might be relatively high in the more favorable situations, and that this enrichment of the country rock may be a condition precedent to a sufficient subsequent concentration to yield workable accumulations.

It is, therefore, inferred that while the processes of sedimentation tended on the whole to leanness, they gave rise to (1) some very important ore-deposits, notably the chief iron ores, the greatest of all ores in quantity and in real industrial value, and (2) a diffuse enrichment of certain other areas which made them productive under subsequent concentrative processes, while the sedimentary formations in general were left barren.

Origin of ore regions.—From these considerations it appears that for the fundamental explanation of "mining regions" we must look mainly (1) to magmatic differentiation, so far as the country rock is igneous, and (2) to sedimentary enrichment, so far as the rock is secondary. The determining conditions in both cases are obscure and unpredictable, but the recognition of such regions, and of the function of preliminary diffuse regional enrichment, contributes to a comprehensive view of the complex processes of ore concentration. The subsequent processes consist in the further concentration of the ore material into sheets, lodes, veins, and similar aggregations by ground-water circulation, or else in the purification of the ores by the removal of useless or deleterious material, or in both combined.

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[1] Chamberlin. Geol. of Wis., Vol. IV, p. 599 et seq., 1882.

Surface residual concentration.—The simplest of all modes of concentration takes place in the formation of mantle-rock. An insoluble or slightly soluble metallic substance sparsely distributed through a rock may be concentrated to working value by the decay and removal of the main rock material, leaving the metallic material in the residuary mantle. The tin ores of the Malay peninsula[1] are especially good examples. The crystals of tin oxide were originally scattered sparsely through granite and limestone, but by their decay and partial removal it has accumulated in workable quantities. Certain gold fields and certain iron ores have acquired higher values in the same way. Such residuary material may be further concentrated by wash into gulches or alluvial flats, in the course of which the lighter parts of the mantle-rock are largely carried away, and the heavier, including the metal or its compounds, are mainly left behind. Gold placers are the best example. The mining of placers by hydraulic processes is but a further extension of the natural process of concentration.

Such concentrates in past ages have in some cases been buried by later deposits, and hence certain ancient sandstones, conglomerates, and mantle-rocks have become ore-bearing horizons. The Rand of South Africa appears to be of this type.

Purification and concentration.—A somewhat different mode of concentration and purification has affected certain of the great iron deposits. As already explained, the iron compounds were originally dissolved from the iron-bearing constituents of the primitive or of igneous rocks, or their derivatives, and were deposited in beds as chemical stratiform deposits. In some cases they were sufficiently pure, as first precipitated, to be worked profitably, but in most cases they were seriously affected by undesirable mineral associates. When, however, such impure deposits are subjected for long periods to the percolation of waters from the surface under favorable conditions, the impurities are often dissolved and the ores concentrated. The great Bessemer ore-deposits of Lake Superior are examples. Originally impure carbonates or silicates, they have been converted into rich and phenomenally pure ferric oxides along certain lines of ground-water circulation, and in certain areas of free leaching. Van Hise has shown the definite relation between the water circulation and the production of the high-grade ores.[1] Vast quantities of unconcentrated lean ores lie in the tracts not thus purified and enriched by circulating waters. This does not appear to be simply residual concentration. The waters seem to have added ferric oxide brought from above, while they carried away the "impurities," silica, carbon dioxide, etc. Perhaps this is an instance of mass action in which the ore present aided in causing additions to itself.

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[1] Penrose. Jour. of Geol., Vol. XI, pp. 135–155, 1903.

Concentration by solution and reprecipitation.—By a process almost the opposite of residual concentration, ore material is often leached out of the surface-rock by water circulating slowly through its pores, cleavage planes, and minute crevices, and is carried on with the circulation until it reaches some substance which causes a reaction that precipitates the ore material. This substance may be a constituent of some rock which the circulating water encounters, such as organic matter. More commonly, the precipitation seems to be due to the mingling of waters charged with different mineral substances, the mingling inducing reaction and the precipitation of the ore. Precipitation, however, does not necessarily follow such commingling. The junctions of underground waterways are sometimes characterized by barrenness instead of richness. In the expressive phraseology of the miners, a tributary current sometimes "makes" and sometimes "cuts out." In chemical phrase, when the mingling waters reduce the solubility of the appropriate substance sufficiently, an ore-deposit is formed; when they increase its solubility, they promote barrenness. Changes of pressure and temperature may enter into the process, and mass action may lend its aid when once a deposit is started.

More concretely stated, the general process of underground ore formation appears to be this: the permeating waters dissolve the ore material disseminated through the rock and carry it thence into the main channels of circulation, usually the fissures, broken tracts, porous belts, or cavernous spaces. If precipitating conditions are found there, deposition takes place. The precipitating conditions may be merely changes of physical state, such as cooling or relief of pressure, but probably much more generally they consist in the commingling and mutual reaction of waters that have pursued different courses and become differently mineralized, as implied above. In these cases the metal-bearing current may be scarcely more important than the precipitating current.

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[1] Van Hise, Mono. XIX, U. S. Geol. Surv., pp. 268–295, 1892.

Since the solvent action is a condition precedent to deposition, the location of the greatest solvent action first invites attention. At present it must be treated in general terms, for it is not known what solutions must be formed beyond the fact that they must include the ore material. Probably they must include much besides. Furthermore, it is not known that deep-seated rocks carry more ore material than similar rocks at or near the surface or at any other horizon. Fantastic conceptions of deep-seated metallic richness are to be shunned as quite beyond practical consideration. The water circulation is probably very slight below a depth of two or three miles at most, and above that depth there is little ground to suppose that the rocks of one horizon are inherently more metalliferous than others of their kind. There is no assignable reason why the igneous rocks at the surface are not as rich in ore material as the igneous rocks two or three miles below, since all are probably eruptive and of much the same nature on the whole, being in many cases parts of the same eruptions.

Location of greatest solvent action.—Solvent action is probably most *intense* where the temperature and pressure are highest, that is, in the deeper reaches of water circulation; but the *amount* of water passing in and out of the deeper zone is but a small fraction of that which courses through the upper horizons, and the total solvent action is quite certainly much greater in the upper zone than in the lower. At the same time the solutions in the upper zone are quite certainly more dilute than those below. The horizon of greatest solution lies between the surface and a level slightly below the ground-water surface, or, in other words, in the zone where atmosphere and hydrosphere cooperate. Surface-waters are charged with atmospheric and organic acids and other solvents, and their general effect upon the rocks is markedly solvent down to or often below the permanent water-level. In this zone concentration by residual accumulation may take place, as already noted, if the metallic compounds resist solution; otherwise this zone is depleted of its ore material by solution, and preparation is made for deposition elsewhere.

Solution also continues to take place varyingly as the water descends below this zone of dominant solution, and extends probably to the full depth of water circulation, but in the deeper circuit, precipitation also takes place and the action becomes complex. With the waters taking up and throwing down material at the same time, it is difficult to estimate the balance of results.

When waters that have been mineralized near the surface descend, they often take on a precipitating phase at no great depth below the upper level of the ground-water; thus sulphides that were oxidized and dissolved near the surface are reprecipitated, often at horizons not greatly below the permanent water-level. Waters that dissolve metallic substances in the upper levels often become charged with sulphuretted hydrogen and other precipitants within a few scores or a few hundreds of feet of the surface, as deep wells abundantly prove. The freshness of surface which metallic sulphides often exhibit at these levels is fair ground for inferring recency of deposition and absence of solvent action. Actual demonstrations of depositions in progress are not wanting.

Short-course action.—The concentration which thus takes place by solution in the upper zone, followed closely by reprecipitation within a few score or a few hundred feet, may well be termed the *short-course mode* of ore concentration. It finds its most important illustration in what is commonly known as the "secondary enrichment" of ore-deposits. The ores in the outcropping edge of the vein or lode are dissolved by the surface-waters, carried a short distance down the ore tract and redeposited, causing enrichment at that point. This is only a special case of what takes place generally at this horizon. It is effective in this case because it has a previous partial concentration to work upon. Secondary enrichments of this kind often contain most or all the workable values of the ore tract. If instead of a previous concentration in a vein, lode, or similar ore tract, there had been partial concentration in the country rock by sedimentation, as in the case of iron-ore beds and perhaps lead-, zinc-, and copper-impregnated sediments, the short-course method may give working values not before possessed. In some of the more obscure cases of previous partial concentration in the crystalline and other rocks, it is probably this short-course action that brings the concentration up to working value. It is probably effective also in concentrating the metallic contents of certain igneous rocks that were rich in metallic material when extruded. How far this is true has been, and still remains, a mooted question.

Long-course action.—After the surface-waters have once passed through a cycle of dissolving and precipitating action, as they are apt to do within the first few hundred feet of their courses below the water-level, they are liable to pass through a succession of dissolving and depositing stages, each reaction resulting in a state that makes a new reaction possible. This is especially true if the waters pursue deep courses. Strictly speaking, the precipitations usually concern only a part of the substances dissolved. New substances are often taken up in the very act of throwing down those already held, and the way thus prepared for further changes. If the water pursues a deep and devious course, it may receive additions by solution and suffer losses by precipitation at many points in its course, both descending and ascending. The changes are very complex, and in the case of a deep or long circuit where various rocks, pressures, and temperatures are encountered, the history becomes one long succession of complexities, the full nature of which is not yet revealed.

In the deeper circuits, each individual current usually takes on a descending, a lateral, and ascending phase, the three being necessary to complete a circuit. The chemical conditions of the waters in the three phases are probably not sharply distinguished from one another, and hence there seems to be no defined horizon of concentration comparable to that near the water-level already described. The chief distinctions in the deeper regions relate to pressure, temperature, length or depth of penetration, and duration of contact. It seems safe to assume, as a general truth, that, other things being equal, the solutions become more complex and more nearly reach general saturation the farther and the deeper the waters penetrate.

It has long been a mooted question whether ore-deposits are due chiefly to descending, to lateral, or to ascending currents. The question in its usual form is too undiscriminating for advantageous discussion, but if the ore-deposits due to surface or short-course concentrations and reconcentrations be set aside, as in some sense a separate class, the relative functions of the descending, the lateral, and the ascending portions of the deeper circulations become a measurably definite question. Two great working factors enter into the comparison: (1) much greater circulation in the upper zone, where lateral movement most prevails; (2) much greater heat and pressure in the lower zone, where the circulation must be chiefly vertical.

Heat and pressure in general favor solution, and hence so far as this factor goes, descending water is likely to be increasing its mineral content, rather than diminishing it by deposition. But this is only general; particular elements of the solution may be deposited. In ascending, as the same water must later, it is predisposed to deposition from loss of solvent power through reduction of pressure and temperature. The theoretical balance is here clearly in favor of preponderant deposition by the ascending portion of the current. So far as precipitation is dependent on the mingling of differently mineralized waters, descending and ascending currents seem to be situated much alike, in general, for both are subject to accessions and mutual unions.

The amount of water that circulates in the deeper horizons is much less than that nearer the surface. Allowing a few hundred, or at most one or two thousand feet for the special short-circuit zone next below the water-level (it is known to reach 1000 to 1500 feet in some cases), the water circulating through the next 1000 or 2000 feet is probably several times greater than all that circulates at greater depths, and this greater circulation above doubtless offsets, in greater or less measure, the intensified action of the deeper circulation. Much of the upper and more rapid circulation is lateral, being actuated by the sloping surface of the ground-water, which in turn is determined by topography, precipitation, and other surface conditions. Theoretical considerations, therefore, favor the view that lateral flow is an important factor in the concentration of ore material. But as descending and lateral currents almost inevitably meet and mingle with ascending currents, it is difficult to distinguish, in the ore-deposits, the special functions of each phase of action. It is even more difficult to determine whether the different phases are not alike essential to the mutual reactions on which the deposition depends. It may be as necessary to have a precipitant as to have a metallic constituent in solution to be precipitated, and what is more, this precipitating agency may be a substance of no economic value in itself and of no obvious relations to the substances that form the ores. If the deposition is due solely to a physical state, as relief of pressure or lowering of temperature, these considerations do not hold.

Summary.—The general results are probably these: In the deeper circuits, more ore material is brought upward and deposited than is carried downward and deposited, so that metallic values are shifted toward accessible horizons. In the lateral currents, more metallic values are shifted toward the trunk-lines of circulation—the great crevices and other waterways—than are carried from these into the rock and distributed, and lateral segregation results. At the same time the atmospheric waters acting at or near the surface concentrate ore values downwards. *The sum total of these processes is to promote the development of the higher ore values in accessible horizons, and along the main lines of circulation.*

The influence of contacts.—As ore-deposits depend on a dissolving state followed by a depositing state of the waters, and perhaps on a complex succession of these alterations, it is obvious that conditions which favor changes of state and the commingling of different kinds of water are apt to be favorable to ore production. At any rate it is observed that many important ore-deposits occur at the contact between formations of different character. The contact of igneous rock with limestone is a rather notable instance. It is not to be inferred that such contacts are generally accompanied by workable ore-deposits, but merely that a notable proportion of workable ore-deposits occur at such junctions. It is rational to suppose that where the chemical nature of the two formations is in contrast, the waters that percolate through the one are likely to be mineralized very differently from those that course through the other, and hence that on mingling at the contact, reactions are specially liable to take place, and that when a valuable metallic substance is present it is liable to be involved and by chance to suffer precipitation. Reactions are the more probable because the contact is likely to be a plane of crustal movement, and hence more or less open and accompanied by fractures, zones of crushed rock and other conditions that facilitate circulation and offer suitable places for ore formation.

The effect of igneous intrusions.—A special case of much importance arises when lavas are intruded into sediments that have previously been partially enriched in the ways above described. The igneous intrusion not only introduces new contact zones, and more or less fracturing, but it brings into play hot waters with their intensified solvent work, their more active circulation, and the reaction between waters of different temperatures. The special efficiency of these agencies is believed to be the determining factor in many cases.

The influence of rock walls.—The rock walls themselves are thought sometimes to be a factor in ore-precipitating reactions. By mass action, they may withdraw a constituent of the solution and destroy its equilibrium in such a way as to cause the precipitation of the metallic constituent. Once deposited on the walls ores aid, by mass action, the further accretion of ores.

The special forms which ores assume in deposition, as beds, veins, lodes, stockworks, disseminations, segregations, etc., are chiefly incidental to the local situation in which the essential chemical or physical change takes place.

CHAPTER VIII.

STRUCTURAL (GEOTECTONIC) GEOLOGY.

The structural phases which rocks assume.—In the previous chapters, the general method by which rocks are formed has been set forth, and many of their structural features have been touched upon incidentally. It remains to assemble the structural features already mentioned, and to consider certain additional structural phases which rocks assume.

STRUCTURAL FEATURES OF SEDIMENTARY ROCKS.

In the deposition of sediments in the sea, or in other bodies of standing water, the coarser portion of the material is usually deposited in the shallow water near the shore where the wave-action is strongest, and the less coarse of various grades is deposited at greater and greater distances from the land, while only extremely fine silt is usually carried out to abysmal depths (see p. 380). To this general law of distribution there are important exceptions. Fine sediments are sometimes deposited near the shore, and where currents, tidal agitation, or floating ice are effective, coarse deposits are occasionally carried far out from the shore.

Stratification.—Sedimentary rocks are usually arranged in more or less distinct layers; that is, they are *stratified*. The stratification consists primarily in the superposition of layers of different constitution or different compactness on one another. Layers of like constitution or compactness are often separated by films of different material which cause the partings between them. The bedded arrangement of stratified rocks is due to various causes, but primarily to the varying agitation of the waters in which the sediment was laid down. Where the depositing waters are agitated to the bottom, coarse sediment is likely to be deposited. Where the waters are quiet at the bottom, fine sediment is the rule. Since the agitation of the waters is subject to frequent change, it follows that coarser material succeeds finer, and finer coarser, in the same place. Hence arise *beds*, *layers*, and *laminæ*. The terms *layer* and *bed* are generally used as synonyms, while *laminæ* are thinner divisions of the same sort. The term stratum is sometimes applied to one layer and sometimes to all the consecutive layers of the same sort of rock. For the latter meaning the term *formation* is often used. Sometimes bedding seems to have been determined by strong currents which temporarily not only prevented deposition over a given area, but even cut away the loose surface of deposits already made, giving a firm surface from which succeeding deposits are distinct. This sequence of events is sometimes shown by the truncation of laminæ, and by other signs of erosion. The commoner sorts of bedded rock are limestones, shales, sandstones, and conglomerates.

The bedding of *limestones* is often caused by the introduction of thin films of clayey material which interrupt the continuity of the lime accumulation and cause natural partings. Sometimes, however, bedding arises from variations in the physical condition of the lime sediment itself. Lamination is not usually conspicuous in pure limestone, though it may be well developed in the shaly phases of this rock. *Shales* are normally laminated as well as bedded, and the lamination is often more notable than the thicker bedding. Bedding in shale may arise from the introduction of sandy laminæ, or by notable changes in the texture of the shale material. Similarly, *sandstones* are sometimes divided into beds by shaly (clayey) partings, but more often by variations in the coarseness of the sand itself, or by the presence of laminæ that are less coherent than those above and below. Sometimes the layers appear to be determined by the compacting of the surface of sand already accumulated before it was buried by later deposits. Sandstones may be thick- or thin-bedded, and their bedding passes insensibly into lamination.

Sand deposits usually take place in relatively shallow water, and the sand is subjected to much shifting before it finds a permanent lodgment. In the course of this shifting, bars are formed which usually have a rather steep face in the direction in which they are being shifted. The sand carried over the top of the bar finds lodgment on the sloping terrace face. The inclined laminæ thus formed constitute a kind of bedding, but since its planes do not conform to the general horizontal attitude of the formation as a whole, it is called *false-* or *cross-bedding* or, more accurately, *cross-lamination* (see Fig. 368). The same structure is developed on delta fronts and generally in water shallow enough to be subject to frequent agitation at the bottom. Sandstone is cross-bedded more commonly than other sorts of sedimentary rock.

The bedding of *conglomerate* is due chiefly to variations in coarseness. Laminæ or thicker layers of sand are frequently found between layers of coarser material. Conglomerate is likely to be thick-bedded, and cross-bedding is common.

Lateral gradation.—When the varying nature of the agitation of the sea at different depths and along the different parts of the coast-border, and during different phases of the sea-currents, is considered, it will be readily understood that sedimentary beds are affected by many irregularities, and that deposits of one kind grade into others horizontally with great freedom. Thus a bed of conglomerate (gravel) may grade

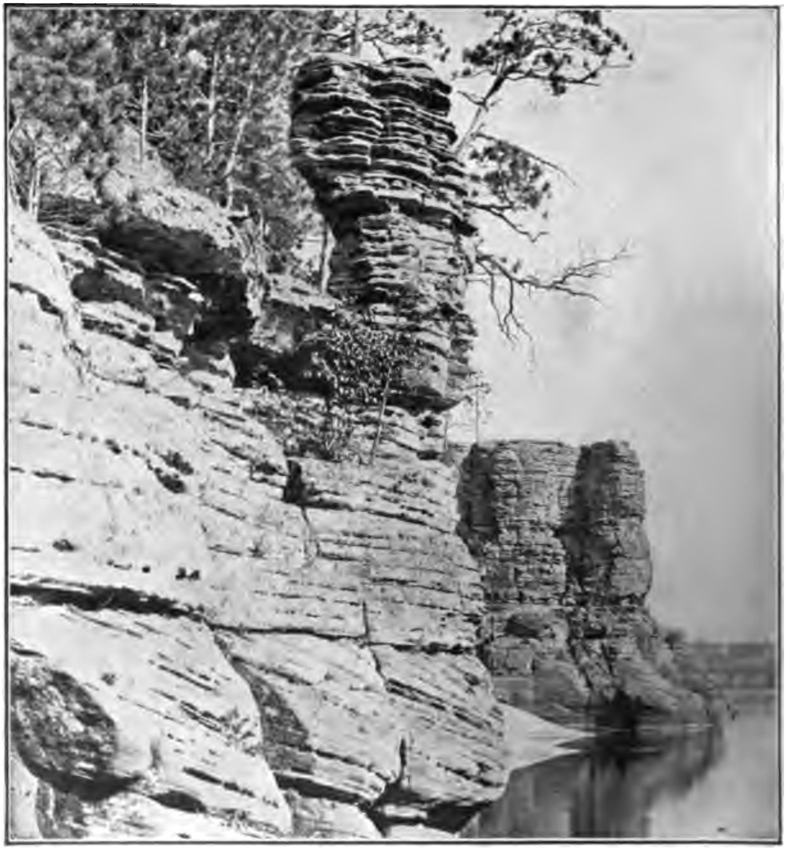

FIG. 368.—Cross-bedding in sandstone. Dells of the Wisconsin near Kilbourn, Wis. (Bennett.)

laterally into sandstone, and this into shale or limestone. It is indeed rather more remarkable that the sedimentary strata should be as regular and persistent as they are, than that they sometimes grade into one another.

Special markings.—The rhythmical action of waves gives rise to undulatory lodgment, known as *ripple-marks* (Fig. 324). They are usually not the direct product of the surface-waves, since they are much too small. They are produced mainly by the vibratory movement of the undertow, but they apparently result from various other phases of vibratory agitation of the bottom waters. They are sometimes made by streams and stream-like currents. Ripple-marks are apparently preserved indefinitely under proper circumstances. They are sometimes found, for example, on very ancient quartzites. Ripples are also made by wind (p. 37). Ripple-marks are usually only an inch or two from crest to crest, but in rare instances they attain much greater size. Examples of ripple-marks 30 feet across are known.[1] Occasional ridges and depressions of much greater dimensions are produced which are attributable to the formation of successive bars, or to the building of wave-cusps.[2] *Rill-marks* are not infrequently produced by the under-

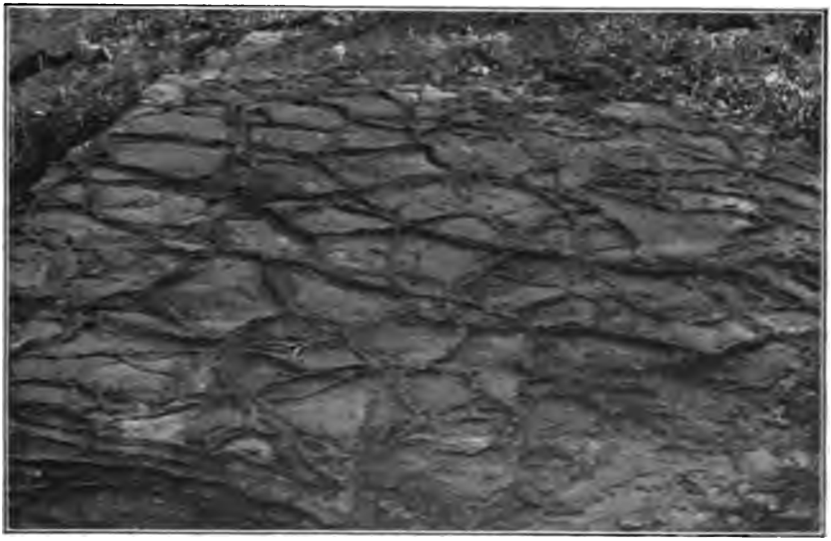

FIG. 369.—Mud-cracks in Brunswick Shale, N. J. (Kümmel.)

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[1] Gilbert. Bull. Geol. Soc. Am., Vol. X, pp. 135–140, 1898.

[2] Branner. Jour. of Geol., Vol. VIII, pp. 481–484, 1900.

tow and other currents passing over pebbles, shells, etc. (Figs. 325 and 326).

Sediments are sometimes exposed between tides, or under other circumstances, for periods long enough to permit drying and cracking at the surface. On the return of the waters, the cracks may be filled and permanently preserved. These are known as *sun-cracks* or *mud-cracks* (Figs. 328 and 369). They chiefly affect shales, but are occasionally seen in limestones and fine-grained sandstones. During the exposure of the sediments a shower may pass and *raindrop impressions* (Fig. 370) be made which are subsequently filled by fine sediment and preserved. The size and depth of rain-drop impressions give some hint as to the meteorological conditions of far-off ages. *Wave-marks*, which consist of the faint line-ridges developed on a sandy beach at the limit of the incoming wave, are sometimes preserved and may be seen occasionally on layers of rock deposited millions of years ago.

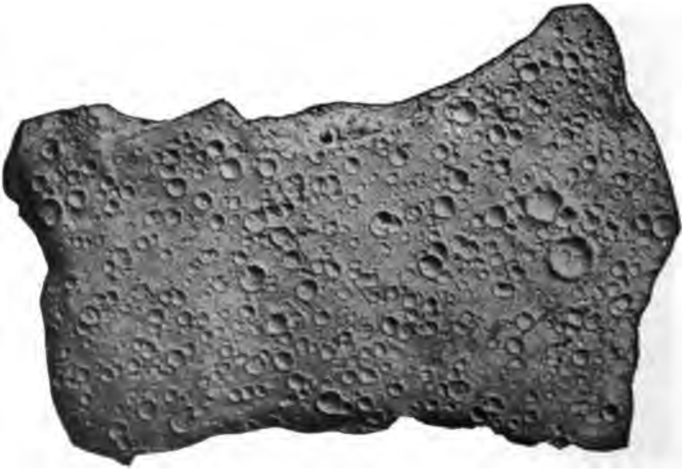

Fig. 370.—Rain-drop impressions. (Brigham.)

Concretionary structure.—Various sedimentary formations contain nodules or irregularly shaped masses of mineral matter unlike the rock in which they occur. When these nodules consist of matter aggregated about some center, they are called *concretions*. They are common in sedimentary rocks, and here it may sometimes be seen that the aggregation has taken place about a shell, a leaf, or some other organic relic. The nuclei are, however, not always organic. The material of the con-

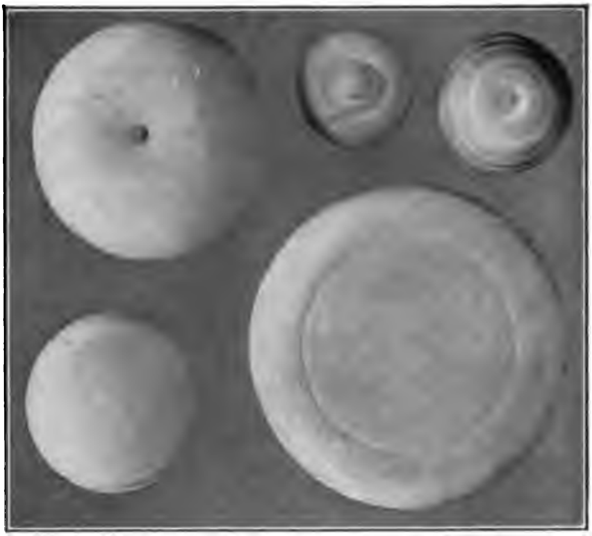

Fig. 371.—Discoid calcareous concretions from post-glacial clays. Ryegate, Vt. (Photo. by Church.)

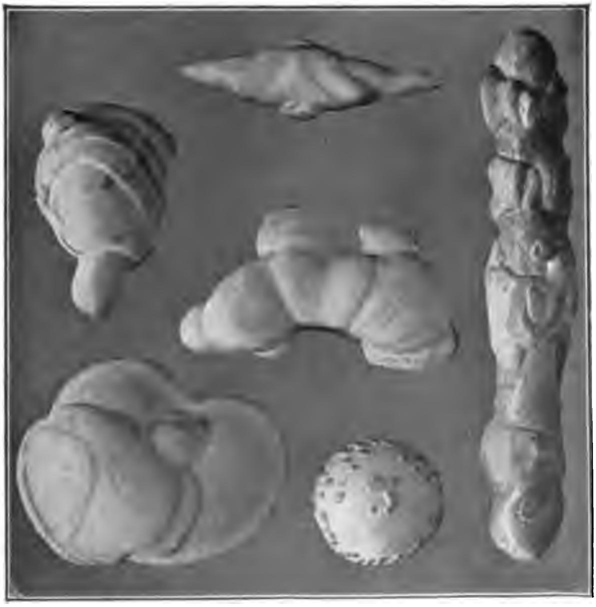

Fig. 372.—Irregular calcareous concretions. Ryegate, Vt. (Photo. by Church.)

cretion may have come from the immediately surrounding rock, having been first dissolved by water and then deposited about the nucleus, or it may have been introduced from without, likewise by the agency of water. In the first case, the mineral matter of the concretion is usually one of the minor constituents of the rock. Thus the commonest concretions in limestone are composed of impure silica (chert, Fig. 361); in shale, of lime carbonate or iron sulphide; in sandstone, of iron oxide. The concretion may be made up almost wholly of concentrated matter,

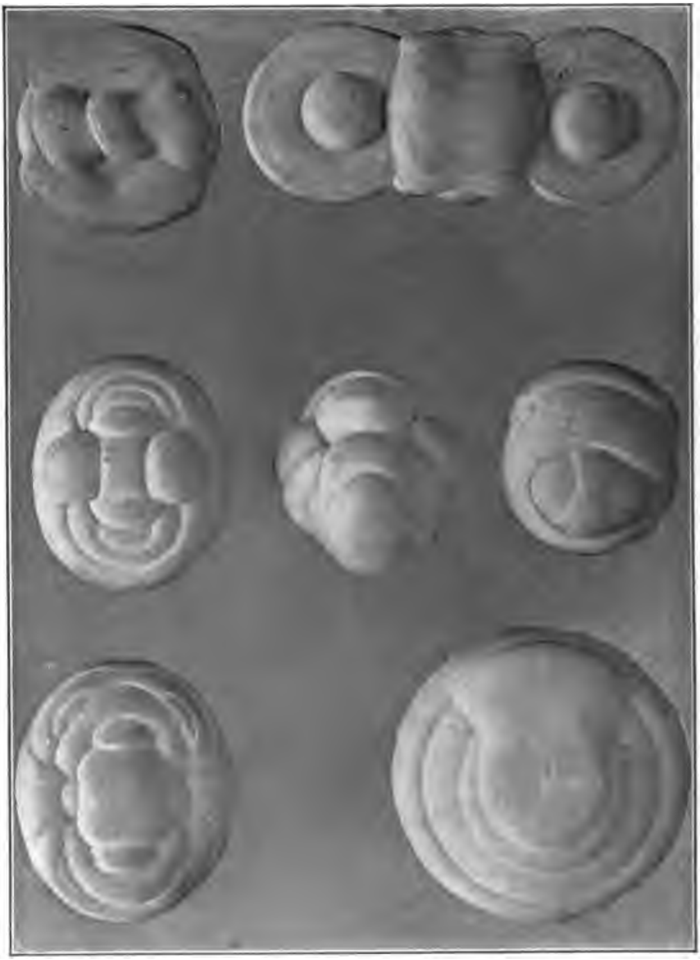

FIG. 373.—Calcareous concretions, some of them showing bilateral symmetry. Ryegate, Vt. (Photo. by Church.)

in which case the matter originally in the place of the concretion has been crowded aside; or it may involve much of the material of the imbedding rock. Thus the concretion of lime carbonate in shale may be nearly pure, or it may involve much of the earthy matter of the shale, while the concretion of iron oxide in sandstone commonly includes much sand. In extreme cases, indeed, the concentrated matter of the concretion merely cements the material involved into distinct nodules. Occasionally the rock substance itself takes on a concretionary form, all or most of its material being involved.

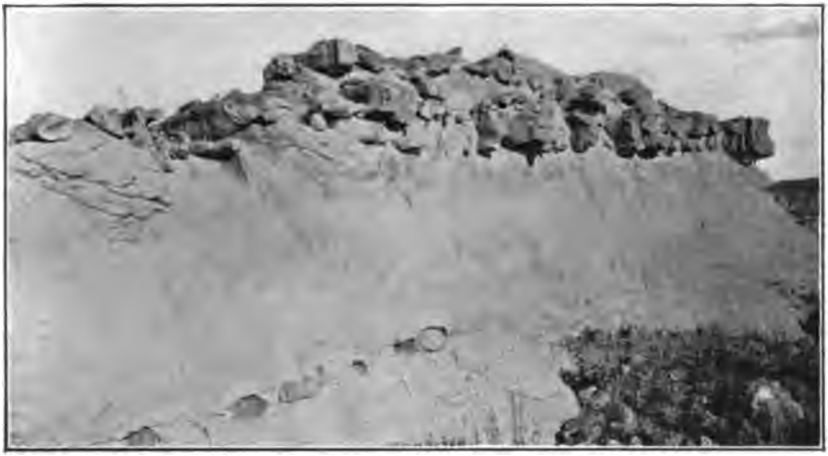

Fig. 374.—Irregular tubular silicious concretions in Arikaree clays. Northwest of Wildcat Mountain, Banner Co., Neb. (Darton, U. S. Geol. Surv.)

In size, concretions may vary from microscopic dimensions to huge masses, 8, 10, or even more feet in diameter. The variations in shape are also great. They may be spherical, elliptical, discoid, or they may assume more irregular and complex forms (Figs. 371 and 372). The conditions of growth have much to do with the form. Thus a concretion which starts as a sphere may find growth easier in one plane than another, when it becomes discoid. Two or more concretions sometimes grow together, giving rise to complicated forms. Some of the most complex and fantastic forms are perhaps to be explained in this way. Concretions sometimes take the form of tubes. Some minute tubular concretions were formed about rootlets, but the larger ones appear to owe their form to other influences (Fig. 374).

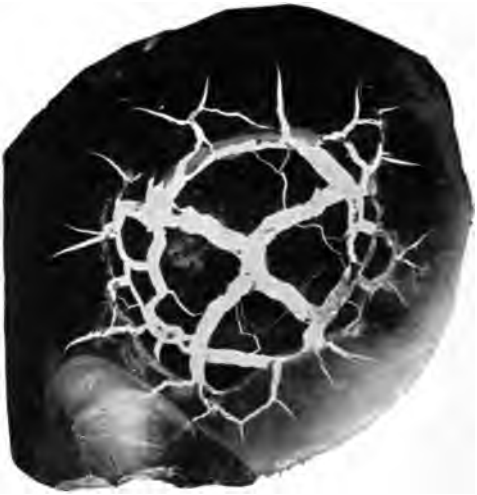

Fig. 375.—Section of a concretion (*septarium*) the cracks of which have been filled by matter deposited from solution. About half natural size. (Photo. by Church.)

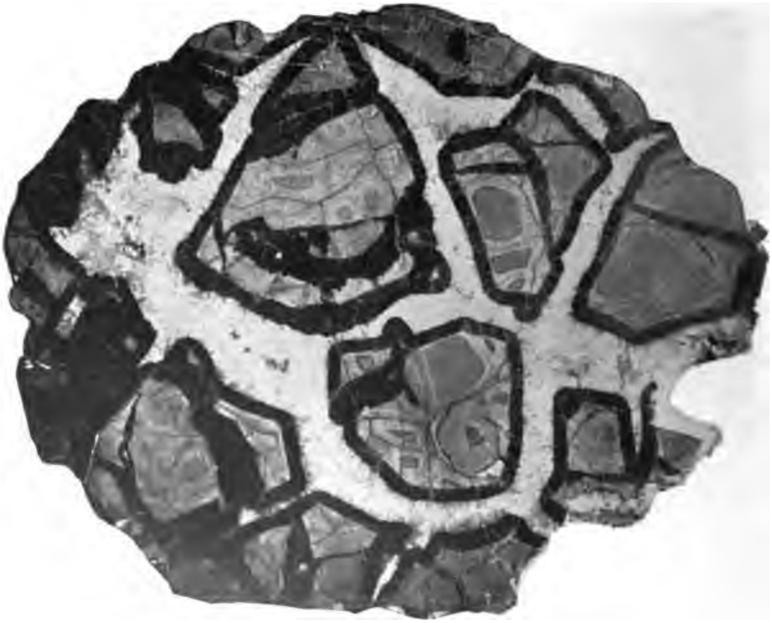

Fig. 376.—Section of a concretion, the cracks in which have been filled by deposition from solution. The filling appears to have wedged the parts of the original concretion apart. The fillings are veins. Some of them show that the vein-material was deposited on both walls. About half natural size. (Photo. by Church.)

One of the most extraordinary features of some concretions of complex form is their symmetry. This may be of various phases; in exceptional cases there is a bilateral symmetry almost as perfect as in the higher types of animals. This is especially true of certain calcareous concretions developed in plastic clays (Fig. 373).

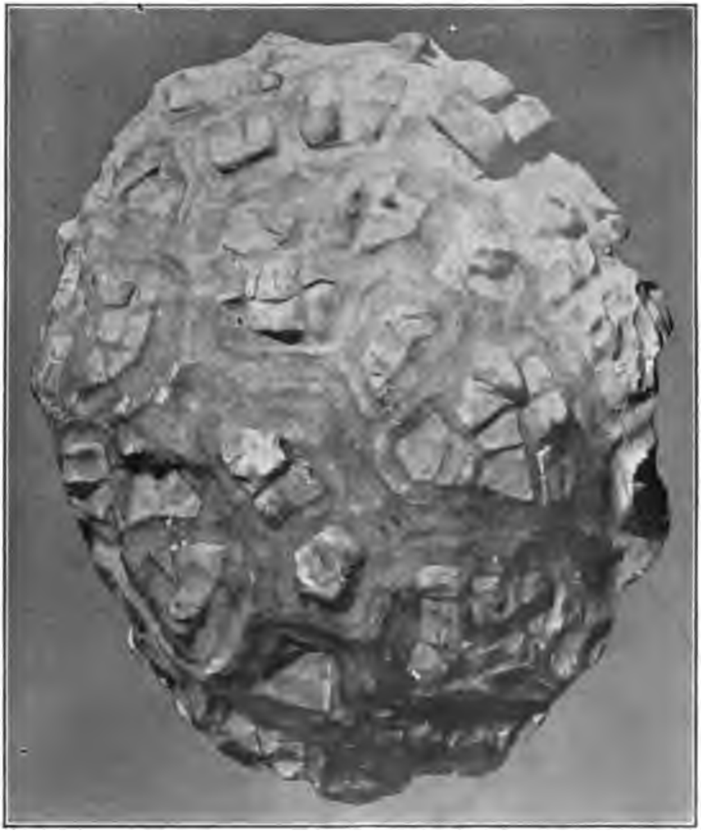

FIG. 377.—Septarium from Cretaceous clays near the east base of the Rocky Mountains in Montana. (Photo. by Church.)

Concretions sometimes develop cracks within themselves, and these may then be filled with mineral matter differing in composition or color from that of the original concretions (Figs. 375 and 376). Concretions the cracks of which have been filled by deposition from solution, are called *septaria*. They are especially abundant in some of the Cretaceous shales and clays. In not a few cases the filling of the cracks appears to have wedged segments of the original concretion farther and farther apart, until the outer surface of the septarium is made up more largely of vein-matter than of the original concretion (Fig. 377). Such concretions are often popularly known as "petrified turtles."

Concretions of the sort indicated above often develop after the enclosing sedimentary rock was deposited. This is shown, among other things, by the fact that numerous planes of lamination may sometimes be traced through the concretions.

Concretions also form in water during the deposition of sedimentary rock. Exceptionally, sedimentary rock is made up chiefly of concretions. The chemical precipitates from the concentrated waters of certain enclosed lakes sometimes take the form of minute spherules which resemble the roe of fish. From this resemblance the resulting rock is called *oolite* (Fig. 357). Oolite is now forming about some coral reefs, presumably from the precipitation of the lime carbonate which was temporarily in solution. Considerable beds of limestone are sometimes oolitic. The calcium carbonate of such rock may be subsequently replaced by silica, so that the oolitic structure is sometimes found in

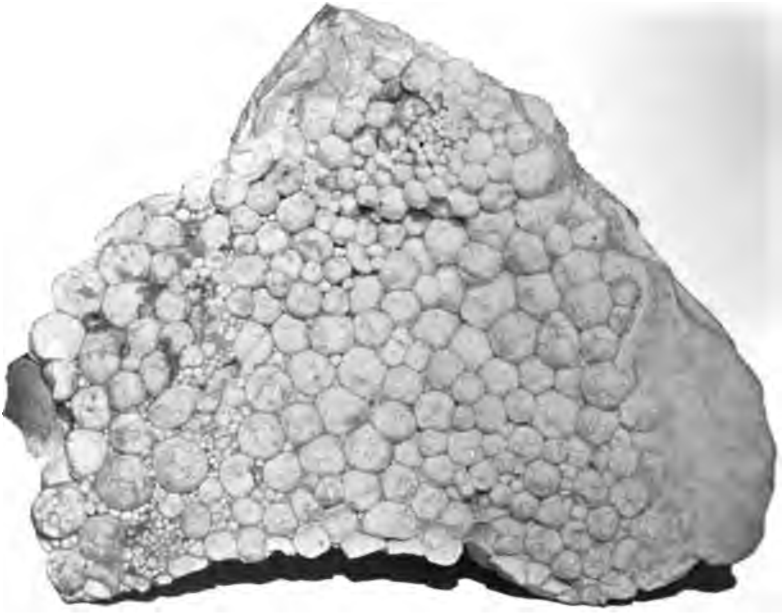

FIG. 378.—Pisolite. Half natural size. (Photo. by Church.)

silicious rock. If the concretions become larger, say as large as peas, the rock is called *pisolite* instead of oolite (Fig. 378).

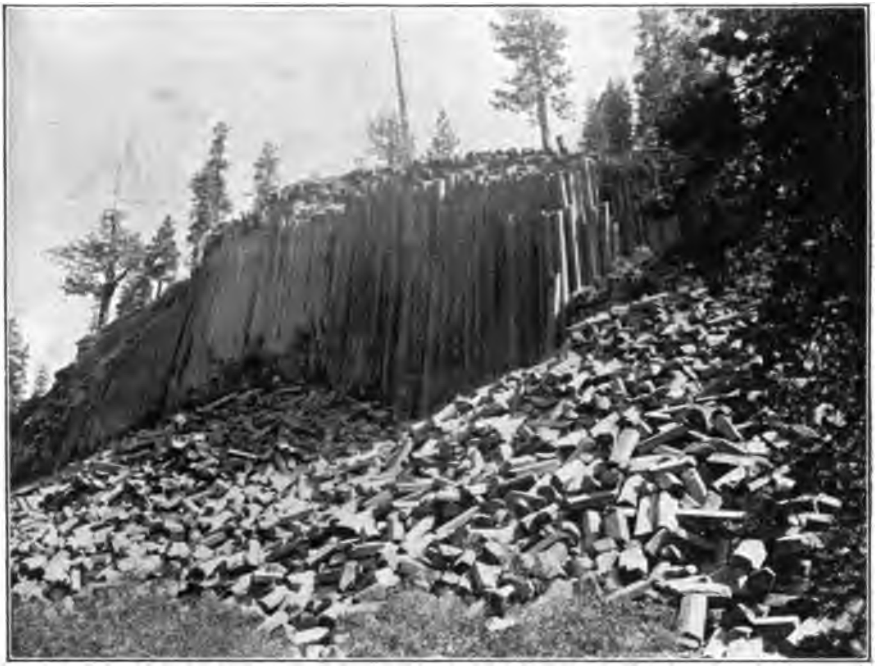

FIG. 379.—Columnar structure, "Devil's Post Pile." Upper San Joaquin Canyon, Sierra Nevada Mountains.

Beds of iron ore are likewise sometimes concretionary. Thus in the Clinton formation there are widespread beds of "flaxseed" ore made up of concretions of iron oxide which, individually, resemble the seed which has given the ore its name. The nucleus in this case is usually too small for identification.

Secretions.—When cavities in rock are filled by material deposited from solution, the result is sometimes called a *secretion*. Secretions therefore grow from without toward a center, while concretions follow the opposite order. Crystal-lined cavities (*geodes*, Fig. 359) and agates (Fig. 358) are examples of secretions. Crystal-lined cavities and veins are the same in principle.

STRUCTURAL FEATURES OF IGNEOUS ROCKS.

Certain structural features of igneous rocks have been mentioned in treating of their origin in the previous chapter. When a great flow of lava spreads out upon the surface, there is no internal lamination or stratification, and the resulting rock is usually classified as massive rather than stratified; but when a succession of flows occur, each individual flow forms a layer, and the series as a whole becomes stratiform.

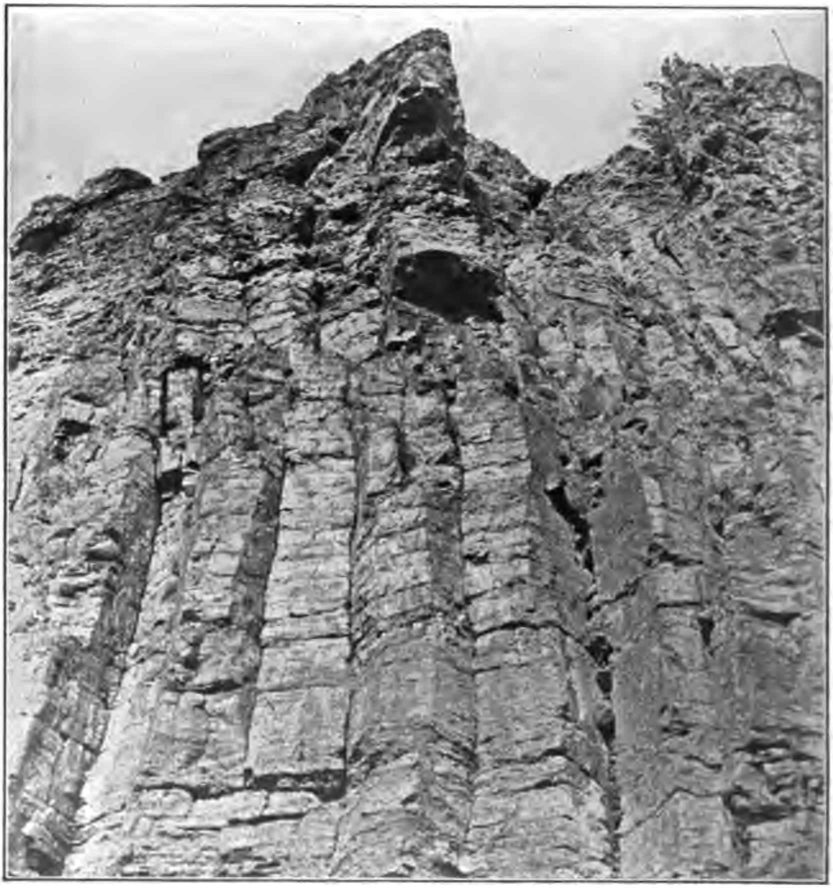

FIG. 380.—Columnar structure, obsidian cliff, Yellowstone Park. (Iddings, U. S. Geol. Surv.)

The successive flows are not usually coextensive. If the later flows of the closing stages of a period of vulcanism fail to reach as far as the earlier ones, a terraced or step-like aspect is given to the region, whence the name *trap-rock* (*trappe*, steps) is derived. Such lava sheets, especially if of basalt, often assume a *columnar* structure in cooling, the columns being rude six-sided prisms standing at right angles to the cooling surfaces (Figs. 379 and 380). This phenomenon is usually best developed where the sheet is intruded between layers of preexisting rock in the form of sills. The formation of the columns is sometimes regarded as a variety of concretionary action, but more commonly as a result of contraction. The former is suggested by the ball-and-socket ends of the sections of some columns (Fig. 382). The development of the columns by contraction may be explained as follows: The surface of the homogeneous lava contracts about equally in all directions on cooling. The contractile force may be thought of as centering about equidistant points. About a given point, the least number of cracks which will relieve the tension in all directions is three (Fig. 383). If these radiate symmetrically from the point, the angle between any two is 120°, the angle of the hexagonal prism. Similar radiating cracks from other centers complete the columns (Fig. 384). A five-sided column would arise from the failure of the cracks to develop about some one of the points (Fig. 385).

When lava is forced into crevices or rises to the surface through fissures, and the residual portion solidifies in them, it gives rise to *dikes*, as illustrated in Figs. 2 and 417 (not a true dike). Dikes are sometimes

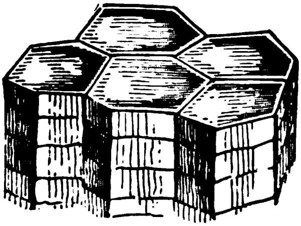
FIG. 381.

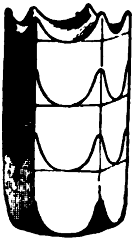
FIG. 382.

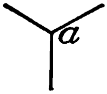

FIG. 383.

FIG. 381.—Sections of columns from Giant's Causeway, coast of Ireland.

FIG. 382.—Ball-and-socket joints in columns of basalt. (Scrope.)

FIG. 383.—Diagram to illustrate the first stages in the formation of hexagonal columns by contraction.

affected by columnar structure. In this case, as in all others, the columns are likely to be at right angles to the cooling surface. Lava solidifying in the passageway leading from the interior of a volcano gives rise to a *neck* or *plug*. If the lava is forced between beds of rock in the form of a sheet, and solidifies there, it is called a *sill*. If, after rising to a certain point in the strata, the lava arches the beds above into a dome, and forms a great lens-like or cistern-like mass, it constitutes a *laccolith* (Fig. 334). If an intrusion of the laccolithic type faults the overlying beds instead of arching them, and especially if the vertical dimension of the intruded mass be great in comparison with its lateral dimensions, its shape is more like that of a plug or core. Such an intruded core is a *bysmalith*[1] (Fig. 124). Between the bysmalith and the laccolith there are various gradations, just as between the laccolith and the sill. When lava forces aside the rocks at considerable depths or absorbs them by solution or by "stoping," and then solidifies in great masses of irregular or undetermined forms, these masses are called *batholiths*.

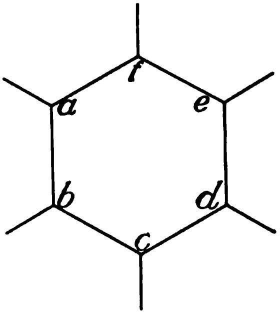

FIG. 384.

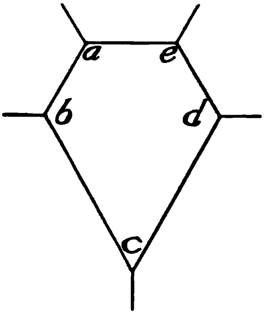

FIG. 385.

FIG. 384.—The completion of the hexagonal columns.
FIG. 385.—Diagram to illustrate the development of five-sided columns.

Volcanic cones are familiar structures built up about the vents of active volcanoes, and will be discussed under vulcanism.

STRUCTURAL FEATURES ARISING FROM DISTURBANCE.

Inclination and folding of strata.—The original attitude of beds, whether formed by water or by lava-flows, is normally horizontal, or nearly so. Both kinds of deposits, however, occasionally take place on considerable slopes. Modifications of the original attitude result from earth movements, and the measurement of these modifications is an important feature of field study. It is recorded in terms of *dip* and *strike*. The dip is the inclination of the beds referred to a horizontal plane, as illustrated in Fig. 386, and is usually measured by a clinometer, the principle of which is shown in Fig. 387. In measuring the dip, the maximum angle is always taken. In Fig. 386, for example, the angle would be less if the direction were either to the right or left of that indicated by the arrow. The direction as well as the amount of the dip is always to be noted. This must be determined by the compass, to which the clinometer may be conveniently attached. Dip 40°, S. 20° W. gives the full record of the position of the bed of rock under consideration. The strike is the direction of the horizontal edge of dipping beds, or more generally, the direction of a horizontal line on the surface of the beds. This is illustrated in Fig. 386. Since the strike is always at right angles to the dip, the strike need not be recorded if the direction of the dip is. Thus dip 40°, S. 20° W. is the same as dip 40°, strike N. 70° W.

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[1] Iddings. Jour. of Geol., Vol. VI, pp. 704–710.

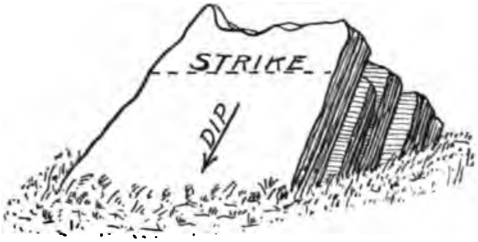


Fig. 386.—Diagram illustrating dip and strike. (Geikie.)

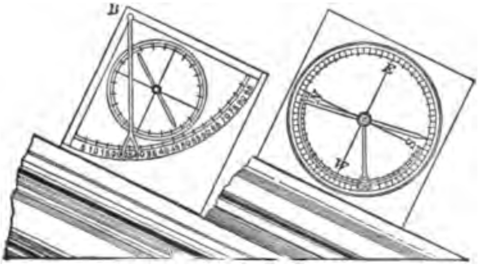


Fig. 387.—The clinometer.

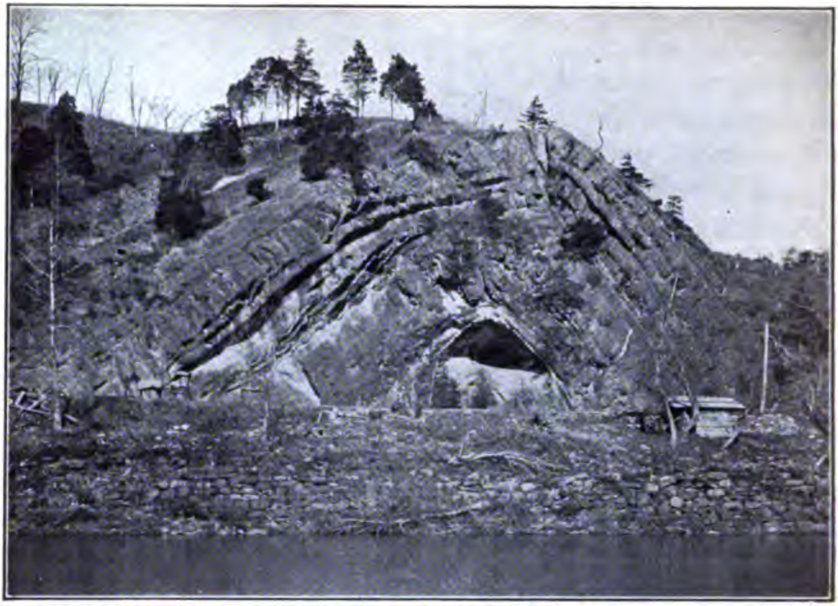


FIG. 388.—Open anticline, near Hancock, Md. (Russell, U. S. Geol. Surv.)

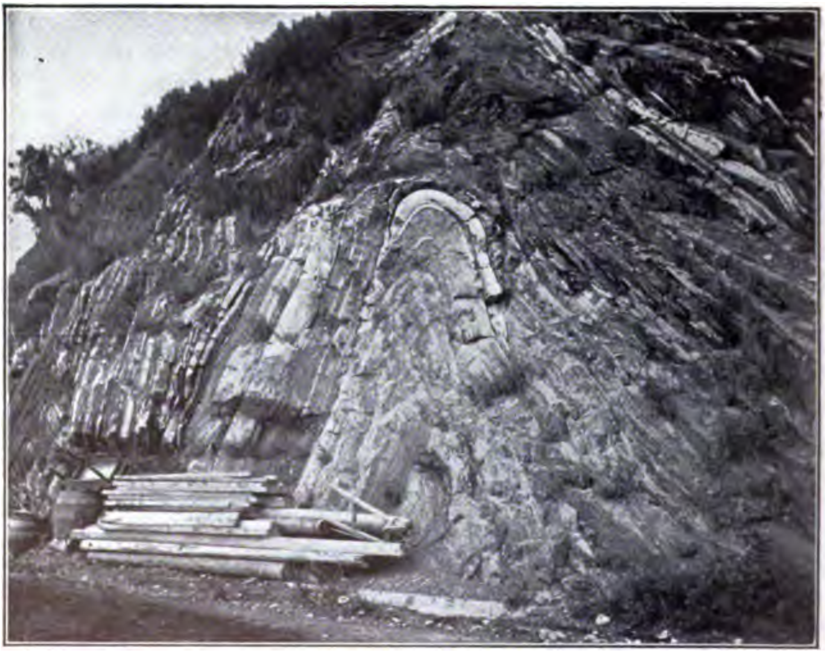


FIG. 389.—Closed anticline, near Levis Station, Quebec. (Walcott, U. S. Geol. Surv.)

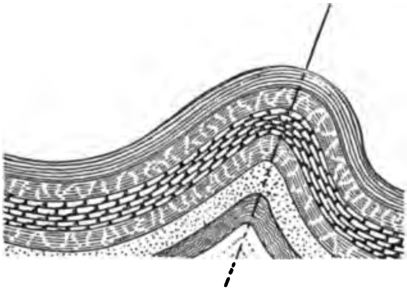


FIG. 390.—Inclined anticline. (Van Hise, U. S. Geol. Surv.)

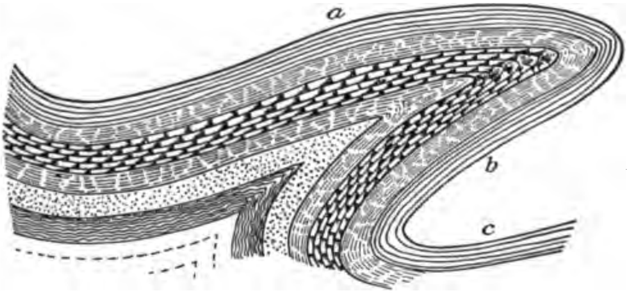


FIG. 391.—Recumbent anticline. (Van Hise, U. S. Geol. Surv.)

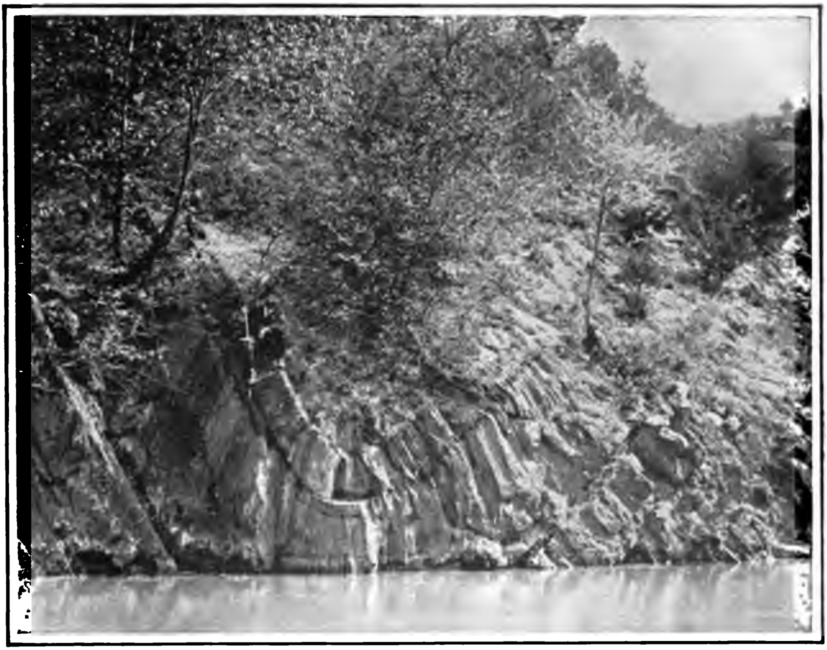


FIG. 392.—Syncline, C. & O. canal, 3 miles west of Hancock, Md. The beds are shale and sandstone near base of the Silurian. (Walcott, U. S. Geol. Surv.)

When the beds incline in a single direction, they form a *monocline*. When beds are arched so as to incline away from one another, they form an up-fold or *anticline* (Figs. 388 to 391). The anticline may depart from its simple form, as shown in Figs. 390 and 391. When beds are curved downward so as to incline towards one another, they form a *syncline* (Fig. 392). When beds assume the position shown in Fig. 393, the folds are said to be *isoclinal*. When they are arched so as to form a cone or dome, and incline in all directions from a central point, they are said to have a *quaquaversal* dip. When considerable tracts are

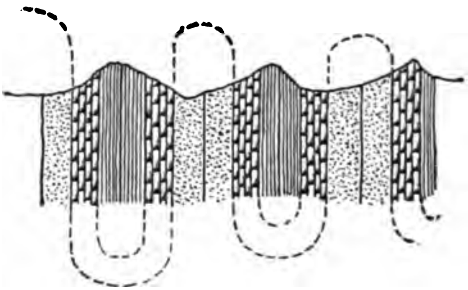


FIG. 393.—Isocline. (Van Hise, U. S. Geol. Surv.)

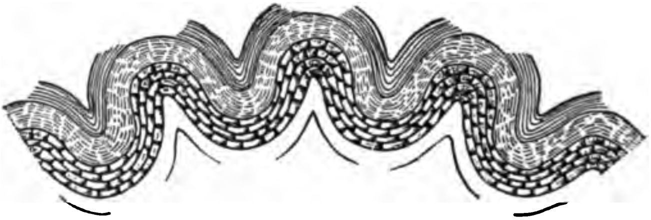


FIG. 394.—Anticlinorium: diagrammatic. (Van Hise, U. S. Geol. Surv.)

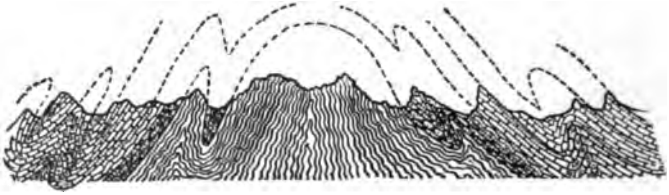


FIG. 395.—Anticlinorium. General section in the central massif of the Alps. (Heim.)

bent so as to form great arches or great troughs with many minor undulations on the flanks of the larger, they are designated as *geanticlines*, or *anticlinoria* (Figs. 394 and 395), and *geosynclines* or *synclinoria* (Figs. 396 and 397). Folding is often accompanied by the development of slaty cleavage (p. 440).

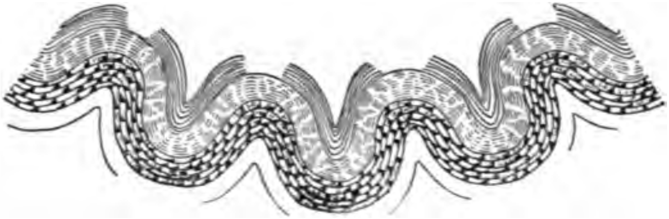


FIG. 396.—Synclinorium: diagrammatic. (Van Hise, U. S. Geol. Surv.)

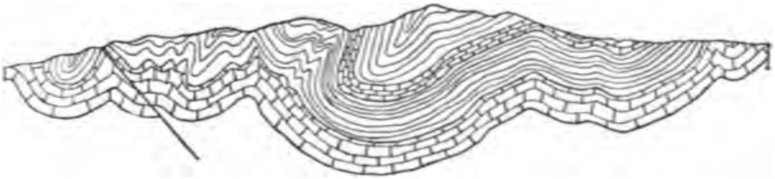


FIG. 397.—Synclinorium, Mt. Greylock, Mass. (Dale, U. S. Geol. Surv.)

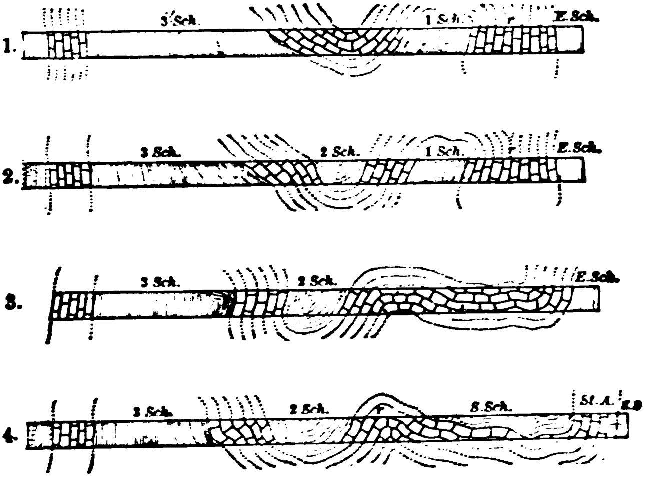


FIG. 398.—A series of diagrams illustrating actual field relations in regions of folded strata. Westchester Co., N. Y. (Dana.)

As found in the field, folds are usually much eroded, and often completely truncated (Fig. 398). The determination of anticlinal or synclinal structure is then not based on topography, or even on such sections as shown in Figs. 394 to 397, for such sections are relatively rare.

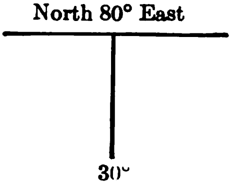


Fig. 399.

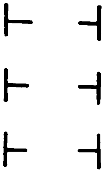


Fig. 400.

Fig. 399.—Diagram to show how dip and strike are recorded.
Fig. 400.—Map record of dip and strike, showing synclinal structure.

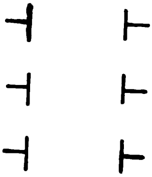


Fig. 401.

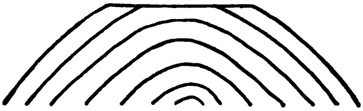


Fig. 402.

Fig. 401.—Map record of dip and strike showing anticlinal structure.
Fig. 402.—The structure of the area shown in Fig. 401, in cross-section.

The structure is determined by a careful record of dips and strikes. On the field map, the record may be made as shown in Figs. 399 to 401, where the free ends of the lines with but one free end point in the direction of dip, while the other lines represent the directions of strike. Applying this method, the structure shown in Fig. 400 represents a syncline, and that in Fig. 401 an anticline. In cross-section, the structure presented by Fig. 401 would appear as in Fig. 402. Fig. 403 shows a doubly plunging anticline; that is, an anticline the axis of which dips down at either end. Fig 404 shows a combination of synclines and anticlines, and Fig. 405 a cross-section along the line *av* of Fig. 404. The outcrops of rock where the dip and strike

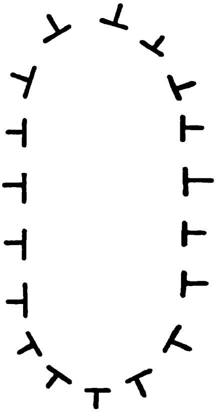
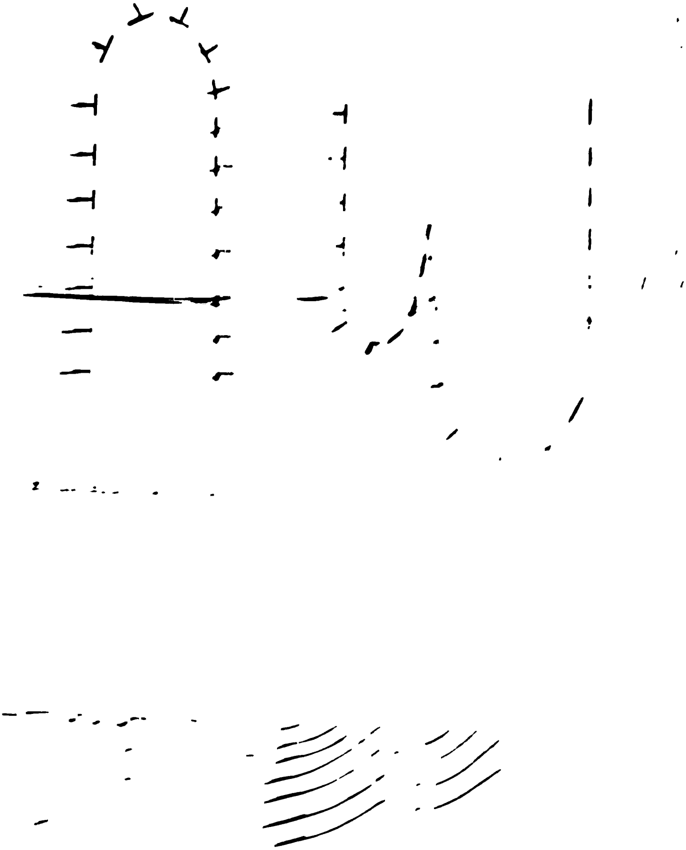


Fig. 403.—Map record of dip and strike showing plunging (dipping down at ends) anticline.



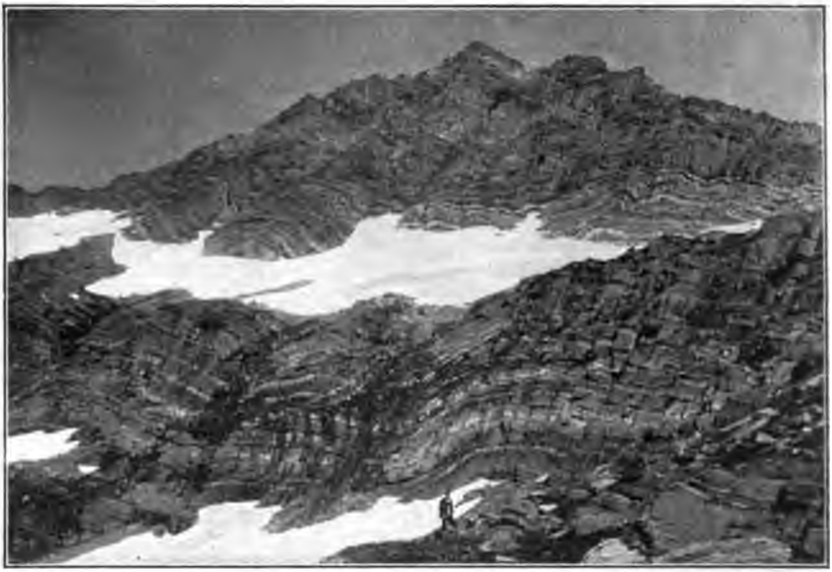


FIG. 408.—Intimate crumpling of beds near head of Sperry glacier, Mont. (Meyers.)

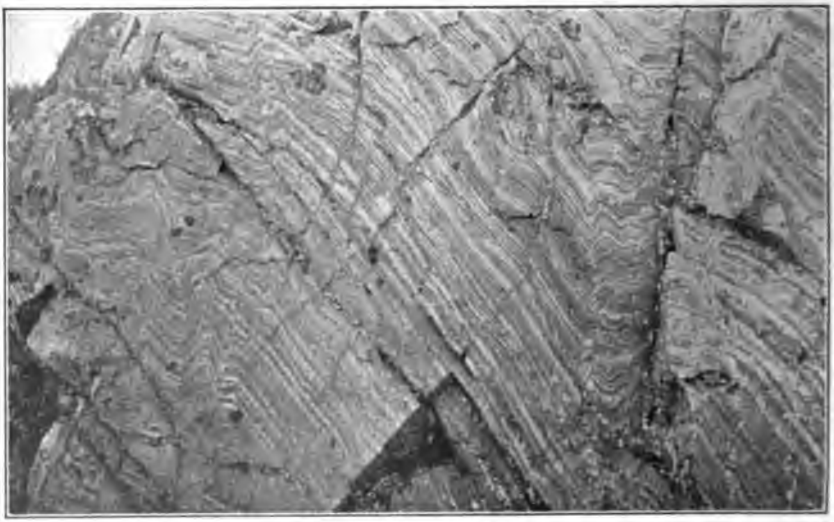


FIG. 409.—Intimate crumpling in detail, accompanied by faulting. Jasper Hill, Ishpeming, Mich. (Meyers.)

may be determined may be few and far between, but when they are sufficiently near one another, the structure of the rock, as shown in Fig. 405, may be worked out, even though the surface be flat.

Much the larger portion of the earth's surface is occupied by beds that retain nearly their original horizontal attitude; but in mountainous regions the beds have usually suffered bending, folding, crumpling, and crushing, in various degrees, in the course of the deformations that gave rise to the mountains. Distortion is on the whole most intense and characteristic in the most ancient rocks known, the Archean, in which a distorted condition is nearly universal, so far as observation goes. Distortion is assigned chiefly to lateral thrust arising from the shrinkage of the earth, as explained in the chapter on Earth Movements.

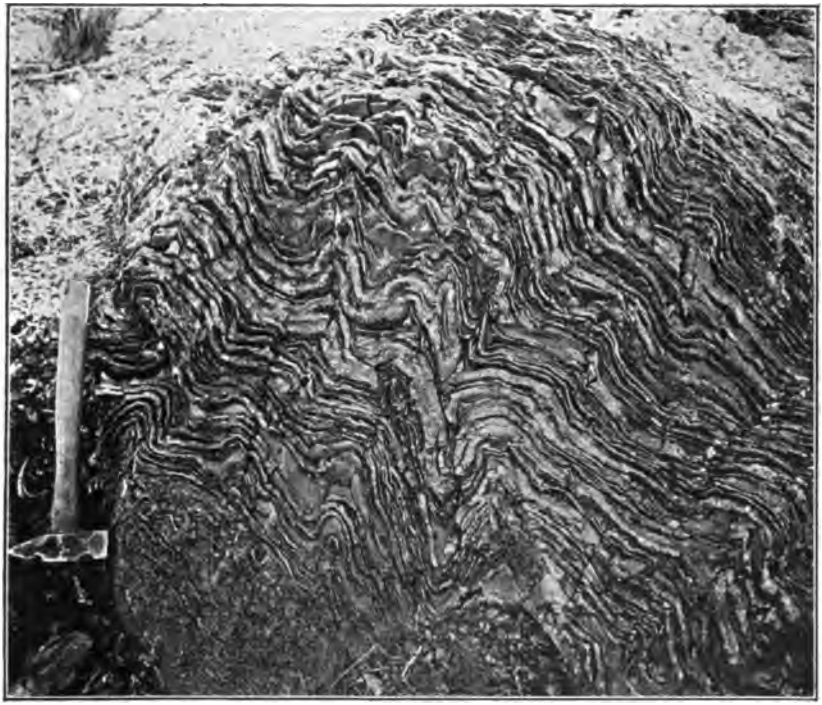


FIG. 410.—Plicated layers of thin-bedded chert in limestone, etched by erosion. Lower Cambrian (?), two miles southwest of Big Pine, Inyo Co., Cal. (Walcott, U. S. Geol. Surv.)

The simpler, and some rather complex forms of deformation, are shown in the preceding figures, but the folding is sometimes much more complex (Fig. 406), the folds sometimes "fan" (Fig. 407), and the beds of which they are composed are sometimes intricately crumpled (Figs. 408 to 410). Among these various phases of deformation there are all gradations and combinations. Overturned folds reverse the order of the strata in the under limb of the fold. After such folds have been greatly eroded, so that their outer form is lost and their relations have become obscure, the reversed beds are likely to be interpreted as though they lay in natural order. In such a case as that represented in Fig. 411, a complex structure may be interpreted as a simple one. Thus the strata of Fig. 411 may have the structure shown in Fig. 412, 413, or 414, so far as dip and strike show.

Joints.—The surface rocks of the earth are almost universally traversed by deep cracks called *joints* (Figs. 415, 138 and 140). In most regions there are at least two systems of joints, the crevices of each system being roughly parallel to one another, while those of the two systems, where there are two, are approximately at right angles. In regions of great disturbance, the number of sets of joints is often three, four, or even more. The joints of each set may be many yards apart, or in exceptional cases, but a few inches, or even a fraction of an inch.

Generally speaking, there are more systems of joints, and more frequent joints in each system, where the rocks are much deformed than where they have been but little disturbed. In undisturbed rocks the joints approach verticality, but in regions where the rocks have been notably deformed, the joint planes may have any position. Not rarely they simulate bedding planes, especially in igneous and metamorphic rocks (Fig. 416). In the latter case especially, the cleavage due to jointing is often mistaken for bedding. They do not ordinarily show themselves at the surface in regions where there is much mantle rock, but they are readily seen in the faces of cliffs, in quarries, and, in general, wherever rock is exposed (Figs. 138 and 140). Though some of them extend to greater depths than rock has ever been penetrated, joints are, after all, superficial phenomena. They must be limited to the zone of fracture, and most of them are probably much more narrowly limited. Joints frequently end at the plane of contact of two sorts of rock. Thus a joint extending down through limestone may end where shale is reached.

Joints are frequently offset at the contact of layers or formations, and a single joint sometimes gives place to many smaller ones. All these phenomena are to be explained on the basis of the different constitution and elasticity of various sorts of rock. Generally speaking, rigid rock is more readily jointed than that which is more yielding.

Joints may remain closed, or they may gap. In the latter case, they may be widened by solution, weathering, etc., but they are quite as likely to be filled by detritus from above, or by material deposited from solution (veins). It is along joint-planes that many rich ore-veins are developed (pp. 478–484).

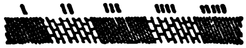


FIG. 411.—This diagram might represent either isoclinal or monoclinal structure. In the former case the strata might have the structure shown in any one of the following Figures, 412 to 414, so far as dip and strike show. (Dana.)

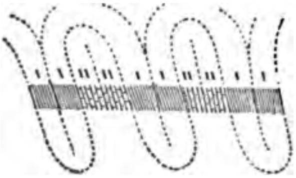


FIG. 412.

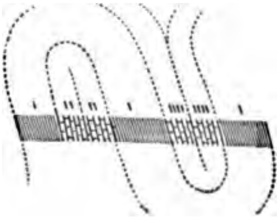


FIG. 413.

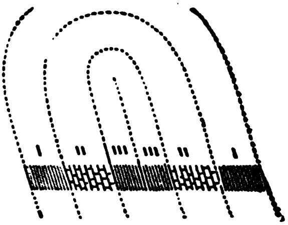


FIG. 414.

FIG. 412.—A possible interpretation of Fig. 411. (Dana.)
FIG. 413.—A possible interpretation of Fig. 411. (Dana.)
FIG. 414.—A possible interpretation of Fig. 411. (Dana.)

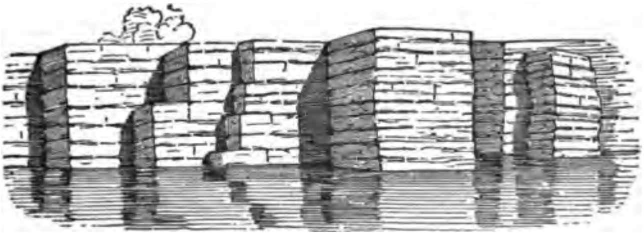


FIG. 415.—Jointed rocks. Cayuga Lake, N. Y. (Hall.)

Joints have been referred to various causes, among which tension, torsion,[1] earthquakes,[2] and shearing[3] are the most important. Most of them may probably be referred to the tension or compression developed during crustal movements.[4] In the formation of a simple fold, for example, tension-joints parallel with the fold will be developed, if tension goes beyond the limit of elasticity of the rock involved. If the axis of a fold is not horizontal, that is, if it "plunges," as it commonly does, a second set of joints roughly perpendicular to the first will be developed. If the uplift be dome-shaped and sufficient to develop joints, they will radiate from the center. It is true that joints affect regions where the rocks have not been folded, and where they have been deformed but little, but deformation to some extent is well-nigh universal.

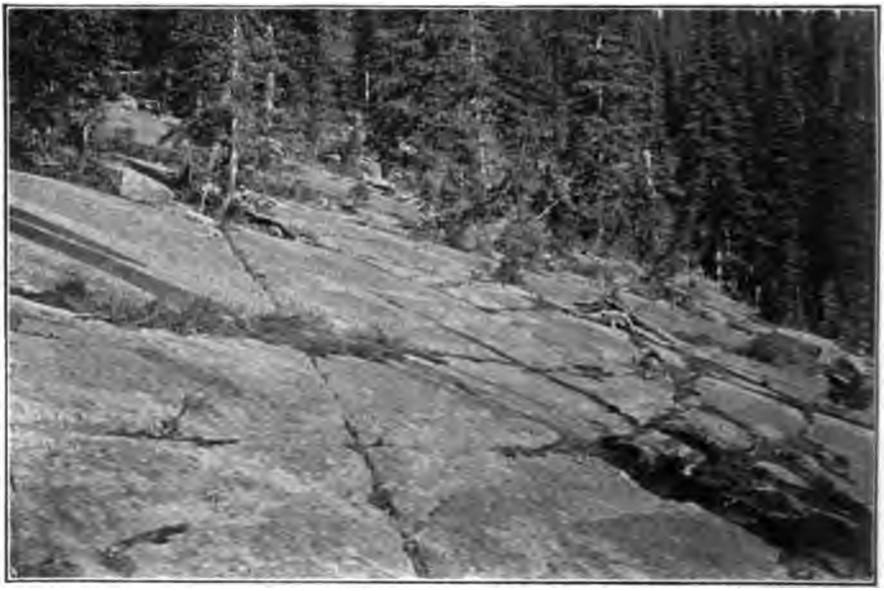


FIG. 416.—Jointing in granite. The surface of the rock is a joint plane. Northwest boundary of the United States. The edges of other joint planes normal to the surface are also shown. (Ransome, U. S. Geol. Surv.)

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[1] Daubrée. Géologie d'Expérimentale, pp. 306–372.

[2] Crosby. American Geologist, Vol. XII, 1893, pp. 368–375.

[3] Becker. Bull. U. S. Geol. Surv., Vol. X, pp. 41–75.

[4] Van Hise. Principles of North American Pre-Cambrian Geology. 16th Ann. Rept. U. S. Geol. Surv., Pt. I, pp. 668–672.

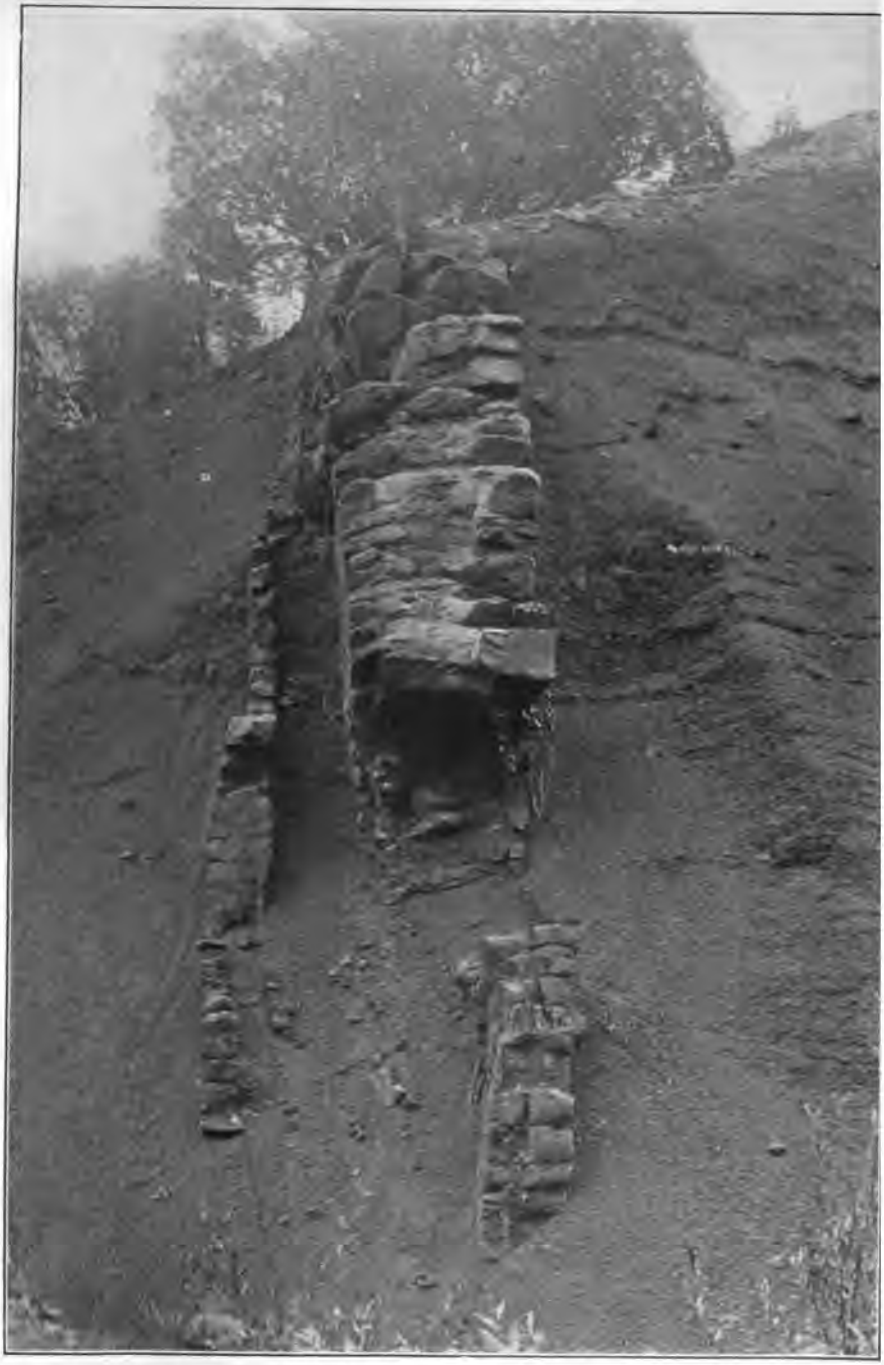

FIG. 417.—Sandstone dike. Northern California. (Diller, U. S. Geol. Surv.)
(Page 513.)

A minor cause of tension-jointing is shrinkage, due (1) to cooling, as in the development of the columnar structure of certain lavas, and (2) to dessication, as shown by the cracks developed in mud when it dries. These causes, however, are not believed to affect rock structures to any considerable depth. Torsional joints and joints due to earthquake vibrations appear to be special phases of tension-joints.

Two or more sets of joints may also be produced by compression, the number being dependent on the complexity of the folding. Many compression-joints correspond in direction with planes of shearing. They are often associated with minor faulting and with slaty cleavage.

Tension-joints appear to be much more widely distributed than compression-joints.

Sandstone dikes.—Exceptionally, open joints are filled by the intrusion of sedimentary material from beneath. Thus have arisen the remarkable sandstone dikes[1] of the West, especially of California (Fig. 417). Such dikes are sometimes several miles (nine at least) in length. The sand of these dikes was forced up from beneath either by earthquake movements or by hydrostatic pressure.

Faults.—The beds on one side of a joint-plane or fissure are sometimes elevated or depressed relative to those on the opposite side; and the displacement is known as a *fault* (Figs. 418 and 419). The joint-planes may have any position, and hence fault-planes may vary from verticality to approximate horizontality. The angle by which the fault-plane departs from a vertical position is known as the *hade* (*bac*, Fig. 418). The vertical displacement (*ac*) is the *throw* and the horizontal displacement (*bc*) the *heave*. The heave and the throw are to be distinguished from the *displacement*, which is the amount of movement along the fault-plane (*ab*, Fig. 418).

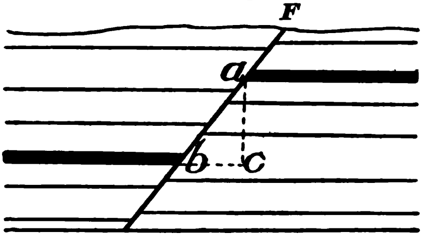

Fig. 418.—Diagram of a normal fault.

The cliff above the edge of the downthrow side is a *fault-scarp*. In many, probably in most cases, the scarp has been destroyed, or at any rate greatly obscured by erosion; but occasionally fault-scarps of mountainous heights, as along the east face of the Sierras and along many of the basin ranges of Utah, Nevada, etc., are found though much modified by erosion (Fig. 419).

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[1] Diller. Bull. Geol. Soc. Am., Vol. I, pp. 441–442. Ibid, Hay, Vol. III, pp. 50–55; and Newsom, ibid Vol. XIV, pp. 227–268.

Faults sometimes arise from over-intense folding (Fig. 420). A deformation which at one point results merely in a bending of the beds, may occasion a fault at another. Faults may pass into folds either vertically (Figs. 421 and 422) or horizontally (Fig. 423). In such cases, thickening and thinning, and stretching and shortening of the beds is often involved (see Figs. 421 and 422). Faults are often due to the greater settling of the beds on one side of a fissure than on the other without special disposition to fold.

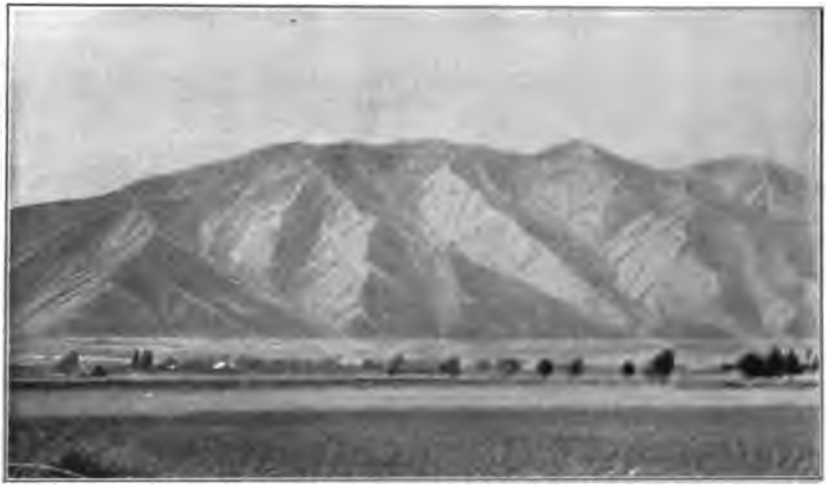

FIG. 419.—A fault-scarp; the triangular faces rising abruptly above the plain at the ends of the spurs. (Davis.)

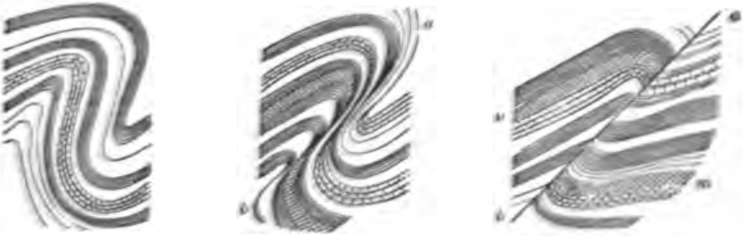

FIG. 420.—Diagrams showing relations of faults and folds.

The rock on either side of a fault-plane is often smoothed

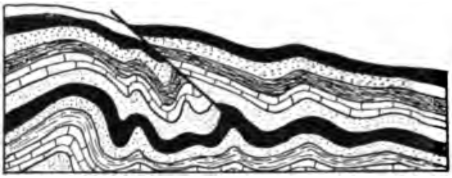

FIG. 421.

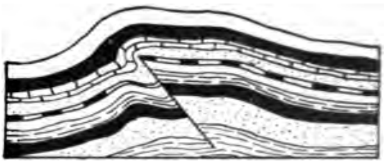

FIG. 422.

FIG. 421.—The fault above grades into a fold below. Thickening and thinning of layers next the fault-plane evident. Based on experimental results of Willis (13th Ann. Rept., U. S. Geol. Surv.)

FIG. 422.—Fault below grading into fold above. Stretching and thinning, and shortening and thickening of beds under pressure is involved. Based on experimental results of Willis.

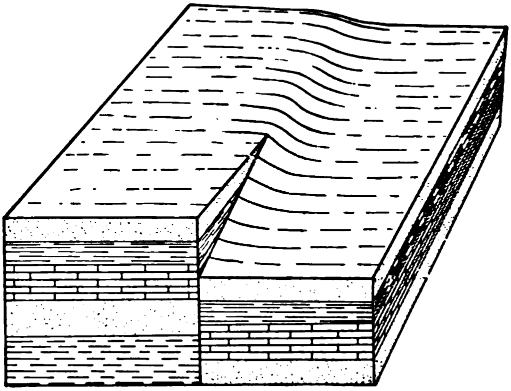

FIG. 423.—Diagram showing a fault grading into a monocline horizontally.

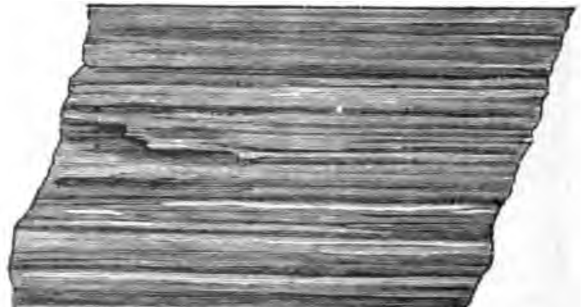

FIG. 424.—Slickenside surface. (Prestwich.)

result of the friction of movement. Such surfaces are *slickensides* (Fig. 424). A slickenside surface has some resemblance to a glaciated surface, but generally gives evidence of greater rigidity between the moving surfaces.

Faults are of two general classes, *normal* and *reversed*. In the normal fault (Fig. 418) the overhanging side is the downthrow side, i.e., the downthrow is on the side towards which the fault-plane inclines, as though the overhanging beds had slidden down the slope. Normal faults, as a rule, indicate an extension of strata, this being necessary to permit the dissevered blocks to settle downwards. In the reversed fault, the overhanging beds appear to have moved up the slope of the fault-plane, as though the displacement took place under lateral pressure. This is clearly shown to be the case where an overfold passes into a reversed fault (Fig. 420). Reversed faults are further illustrated by Figs. 425, 426, and 427. Where the plane of the reversed fault approaches horizontality, the fault is often called a *thrust-fault*, or an

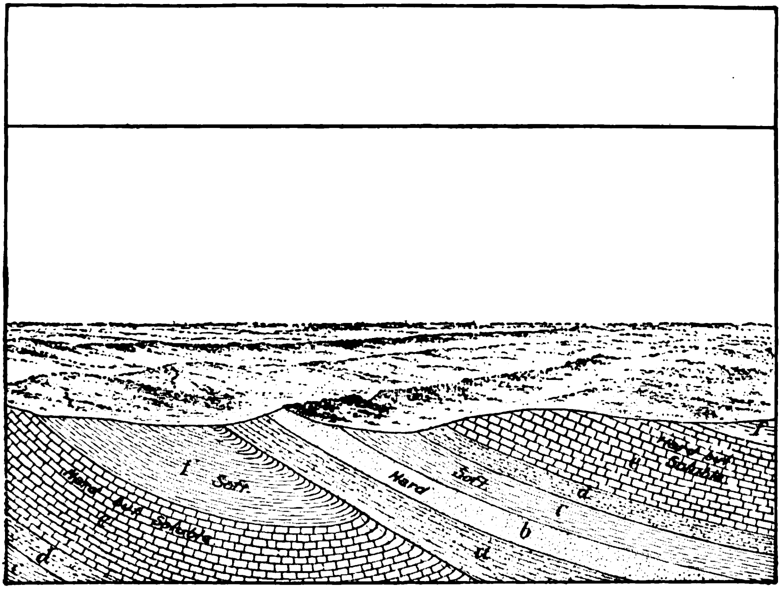

Fig. 425.—Perspective view and vertical section of a thrust-fault. (Willis, U. S. Geol. Surv.)

overthrust. In such cases the throw is to be distinguished from the *stratigraphic throw* (see Fig. 426). In thrust-faults, the heave is often great. The eastern face of the Rocky Mountains near the boundary-line between the United States and Canada has been pushed over the strata of the bordering plains to a distance of at least eight miles.[1] Overthrusts of like gigantic displacement have been detected in British Columbia,[2] Scotland,[3] and elsewhere.

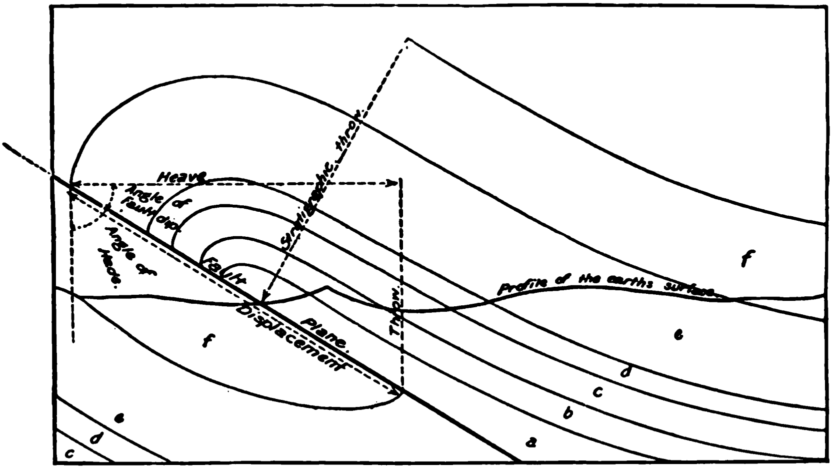

FIG. 426.—Diagram of a thrust-fault illustrating the several terms used in describing faults. The distinctions between heave and displacement, and between throw and stratigraphic throw, are to be especially noted. (Willis, U. S. Geol. Surv.)

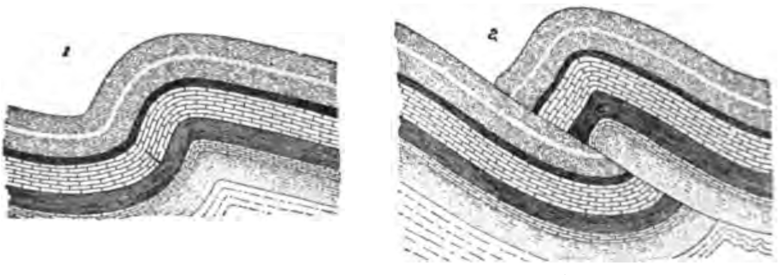

FIG. 427.—Step-fold showing (in 1) break in the massive limestone bed which determines the plane of the break-thrust along which the displacement shown in 2 takes place. (Willis, U. S. Geol. Surv.)

Sometimes a fault branches (Fig. 428) and sometimes the faulting is distributed among a series of parallel planes at short distances from one another,[1] instead of being concentrated along a single plane, thus giving rise to a *distributive fault* (Fig. 429). This is perhaps more common in normal than in reversed faulting.

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[1] Willis. Bull. Geol. Soc. of Am., Vol. XIII, pp. 331–336.

[2] McConnell. Canada Geol. and Nat. Hist. Surv., 1886, Pt. II.

[3] Geikie. Text-book of Geology.

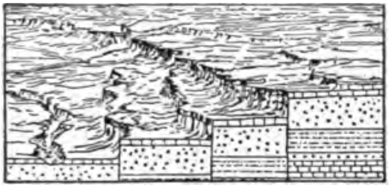

FIG. 428.

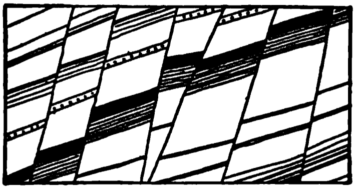

FIG. 429.

FIG. 428.—Branching-fault. (Powell.)
FIG. 429.—Diagram showing a series of small faults—distributive faulting.

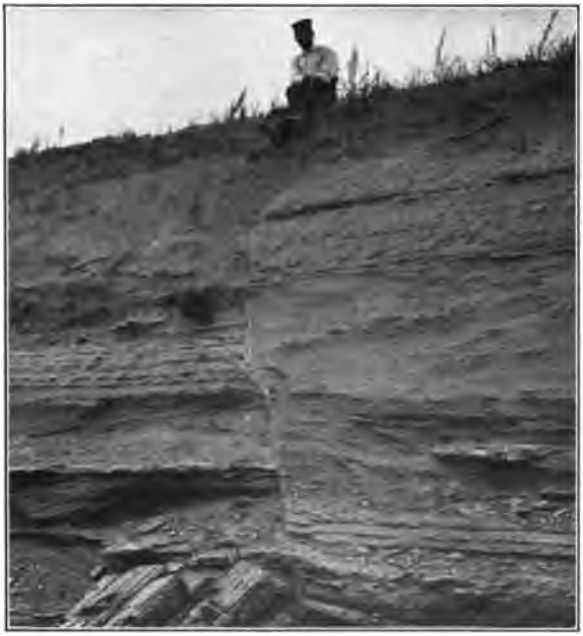

FIG. 430.—Fault in Gering series. Near Rutland Siding, near Crawford, Neb. (Darton, U. S. Geol. Surv.)

The amount of throw occasionally reaches several thousand feet. Occasionally faults of incredible dimensions are reported, but these are perhaps misinterpretations. Faults are observed to die out gradu-

[1] Becker. Geology of the Comstock Lode, Mono. III, U. S. Geol. Surv., Chapter IV.

ally when traced horizontally, sometimes by passing into monoclinal folds, and sometimes without connection with folding. In depth they probably die out in similar ways in most cases. Where the throw is

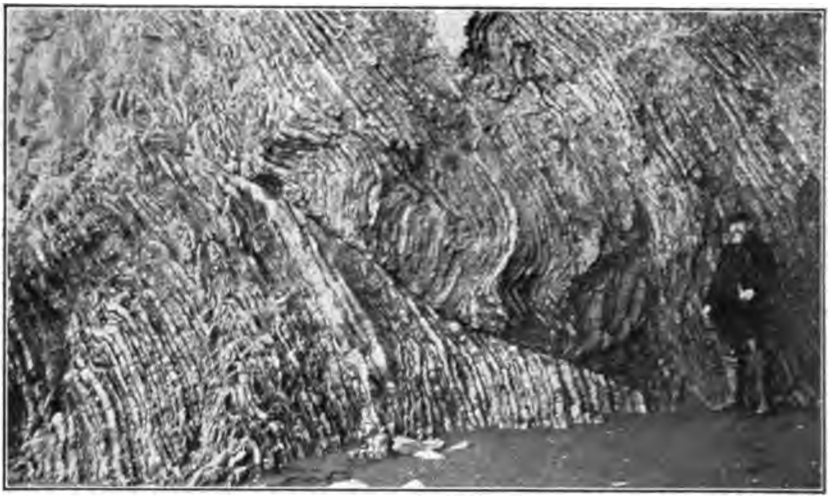

FIG. 431.—Contorted and faulted laminated rock. Cook Inlet. (Gilbert, U. S. Geol. Surv.)

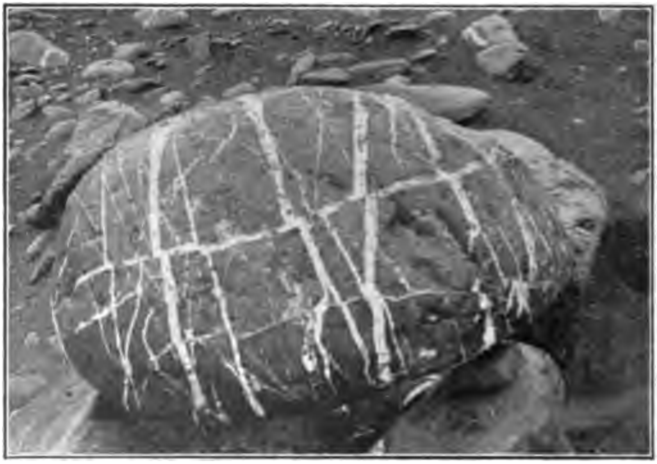

FIG. 432.—Faulting shown in a cobblestone. The fault-planes have become veins by deposition from solution. The figure shows how the relative ages of crossing-faults may be determined. (Schrader, U. S. Geol. Surv.)

great, they probably give place to folds below (Fig. 421). Other phenomena of faulting are illustrated by Figs. 430–435. A fault of thousands, or even hundreds of feet is probably the sum of numerous smaller slippings distributed through long intervals of time. Faulting is probably one of the common causes of earthquakes.

The significance of faults.[1]—Faults afford a valuable indication of the conditions of stress to which a region has been subjected, but some caution must be exercised in their interpretation. Normal faults usually indicate an extension of the surface sufficient to permit the fault-blocks to settle down unequally. Reversed faults usually signify a compression of the surface which requires the blocks to overlap one another more than they did before the faulting. In other words, normal faulting usually implies tensional stress, and reversed faulting compressional stress. It is not difficult to see, however, that in an intensely compressed and folded region there might be cases of normal faulting

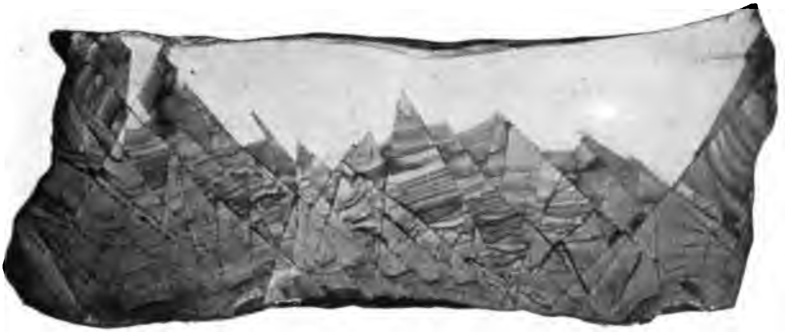

FIG. 433.—Figure showing minute faulting. The length of the specimen is 8 inches. The number of faults is nearly 100. (Photo. by Church.)

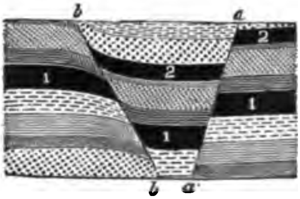

FIG. 434.

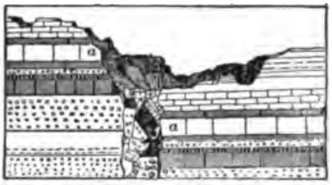

FIG. 435.

FIG. 434.—Diagram illustrating common phenomena of a faulted region. (Dana.)

FIG. 435.—Diagram showing a fault, the plane of which forms an open fissure and has been filled with débris from above. (Powell.)

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[1] Reference, Van Hise. Sixteenth Ann. Rept. U. S. Geol. Surv., Pt. I, pp. 672–678.

on the crests of folds where local stretching took place, and that reversed faults might occur even in regions of tension. But such cases must usually be local, and capable of detection and elimination by a study of the phenomena of the surrounding region. These exceptional cases aside, the general inference from prevailing normal faults is that the regions where they occur have undergone stretching, while the inference from the less widely distributed reversed faults is that the surface where they occur has undergone compression.

In view of the current opinion that the crust of the earth has been subjected to great lateral thrust as a result of cooling, it is well to make especial note of the fact that the faults which imply *stretching* are called *normal* because they are the more abundant; and that the faults which imply thrust are less common, and are styled *reversed*. The numerical ratio of normal to reversed faults has never been closely determined, but normal faults very greatly preponderate, and are estimated by some writers to embrace 90-odd per cent. of the whole. The testimony of normal faults is supported by the prevalence of gaping crevices, and of veins which are but crevices that stood open until they were filled by deposition. All these phenomena seem to testify to a *stretched* condition of the larger part of the surface of the continents. This will again claim attention in the study of Earth Movements.

Effect of faulting on outcrops.—Faulting may bring about numerous complications in the outcrop of rock formations. In a series of formations having a monoclinal structure (Fig. 436), many changes may be introduced. Let it be supposed in the following cases that, after faulting, the surface has been reduced to planeness by erosion. If the fault-plane be parallel to the strike of the beds (*ab*, Fig. 436), and hence a *strike fault*, the outcrop of a given layer may be duplicated (*H*, Fig. 437), or it may be eliminated altogether (Fig. 438). If the fault-plane be parallel to the direction of dip (*cd*, Fig. 436), a *dip fault*, the layer *H* will outcrop, as in Fig. 439, if the downthrow was on the far side, or as in Fig. 440 if the downthrow was on the opposite side. In both cases the outcrop *H* is *offset*, the amount of the offset decreasing with increasing angle of dip and increasing with increasing throw of the fault. If the fault

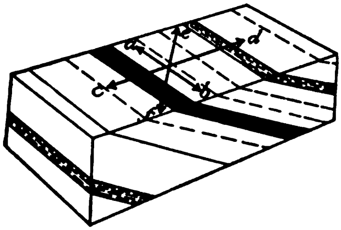

FIG. 436.—Diagram showing an area of rock with monoclinal structure. One layer notably unlike the others.

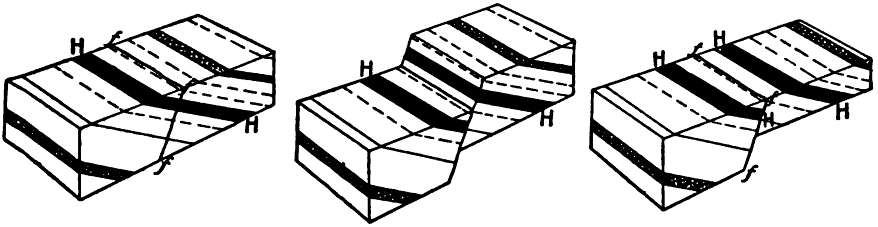

Fig. 437.—Same as Fig. 436, after (1) displacement by a strike fault and (2) base-leveling. The outcrops of certain beds are repeated.

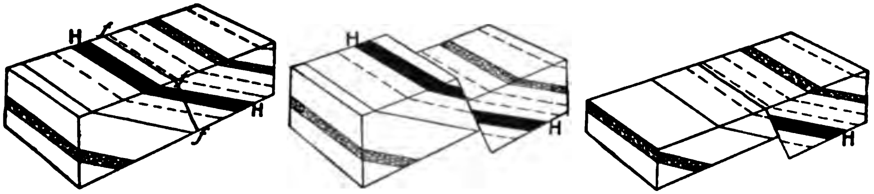

Fig. 438.—Diagram illustrating how a strike fault in such a structure as that shown in Fig. 436 may cause the outcrop of certain beds to disappear.

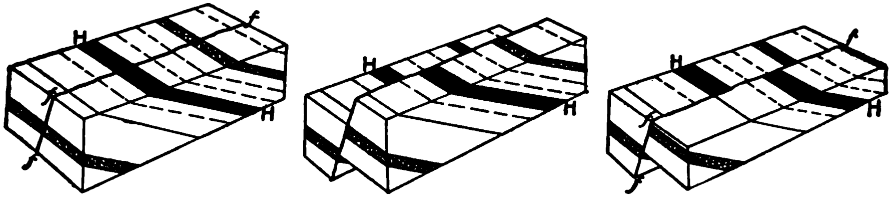

Fig. 439.—Diagram illustrating how a dip fault in the structure shown in Fig. 436 affects the outcrop when the downthrow was on the farther side of the fault-plane.

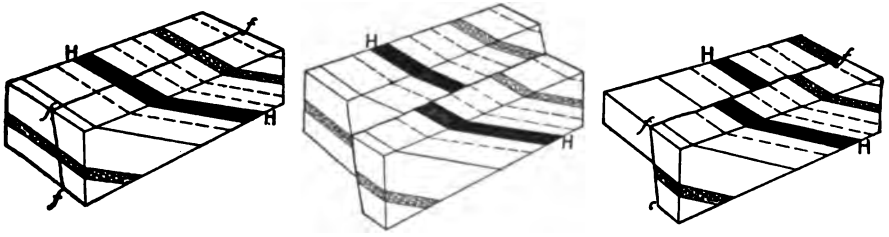

Fig. 440.—Same as Fig. 439, except that the downthrow was on the opposite side.

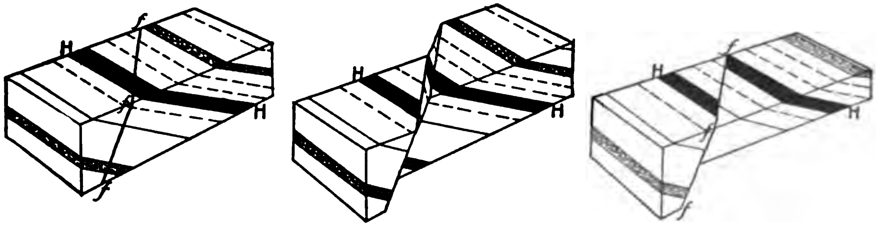

Fig. 441.—Oblique fault in the structure shown in Fig. 436. The downthrow was on the left side. The outcrop of layer *H* is offset with *overlap*.

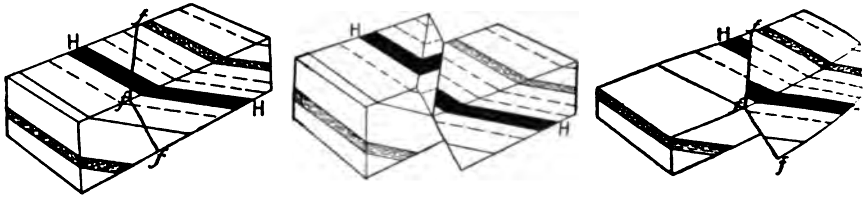

FIG. 442.—Same as Fig. 441, except that the downthrow was on the right side, and the offset is with a *gap* instead of an overlap.

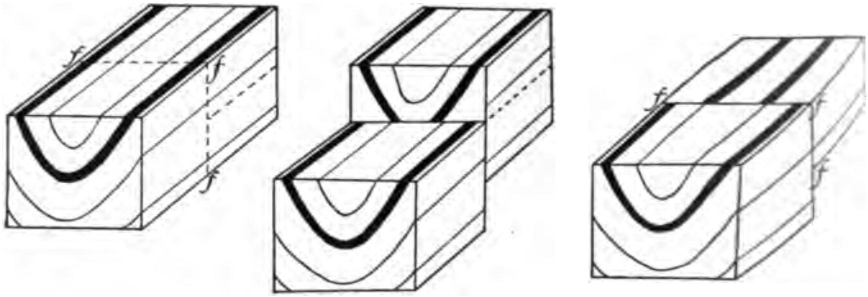

FIG. 443.—Diagram showing effect of faulting on the outcrops of synclinal beds.

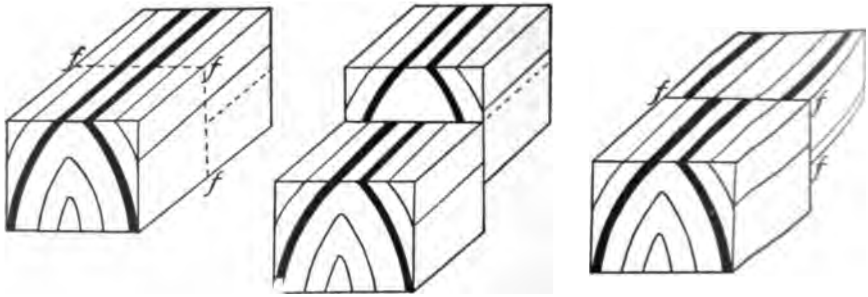

FIG. 444.—Diagram showing effect of faulting on outcrops of anticlinal beds.

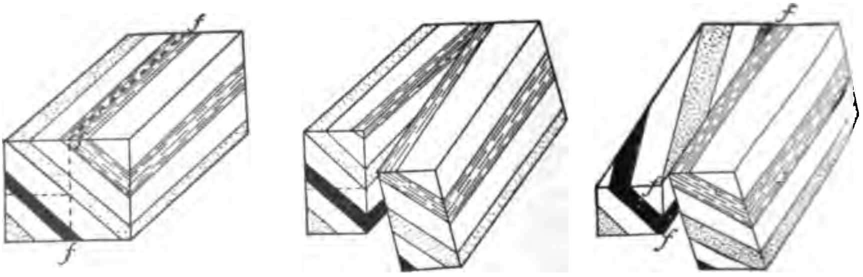

FIG. 445.—Diagram illustrating the effect of diminishing throw on outcrops in regions of folded rocks.

be oblique to the direction of dip and strike (*ef*, Fig. 436), an *oblique fault*, the outcrop of such a layer as *H* will have the relations shown in Fig. 441 if the downthrow was to the left, and that shown in Fig. 442 if the downthrow was to the right. In the former case, it is said that there is *offset with overlap;* in the latter, *offset with gap*. The amount of the overlap and gap, respectively, increases with the increase of throw and hade, and decreases with increase of dip. In all cases the outcrop (after the degradation of the upthrow side) is shifted down dip.

If a fault crosses folds at right angles to their axes, the effect is to change the distance between the outcrops of a given bed on opposite sides of the fault, after the truncation of the folded beds. The distance is decreased on the upthrow side of a syncline (Fig. 443) and increased on the upthrow side of an anticline (Fig. 444). If the throw of a fault in tilted beds diminishes in one direction, it may cause beds to outcrop, as shown in Fig. 445. Various other complications arise under other circumstances. Since faults rarely show themselves in the topography of the surface, except under special circumstances (see p. 151), their detection and measurement is usually based on the study of the relations of the beds involved, as illustrated by Figs. 436–445.

CHAPTER IX.

THE MOVEMENTS AND DEFORMATIONS OF THE EARTH'S BODY (DIASTROPHISM).

THE body of the earth is subject to an infinite variety of movements, ranging from the almost inconceivably rapid to the almost imperceptibly slow, and from the almost immeasurably minute to the enormously massive; but, for practical treatment, they fall mainly into two couplets: (1) the minute and rapid, and (2) the slow and massive. Sudden movements of local masses, giving rise to intense vibrations, are put in the first class. There are innumerable minute and slow movements, but unless they rise to appreciable magnitude by long continuance, they are neglected.

MINUTE AND RAPID MOVEMENTS.

The crust of the earth is in a state of perpetual tremor. For the most part, these tremors are too minute to be sensible, but are revealed by delicate instrumental devices. Some of them are but the declining stages of sensible vibrations, but others are minute from their inception. Many of them spring from the ordinary incidents of the surface, and claim attention chiefly as obstacles to the study of more significant oscillations. Winds, waves, waterfalls, the tread of animals, the rumble of traffic, the blasts of mines, the changes of temperature, the variations in atmospheric pressure, the weighting of rainfall and the lightening of evaporation, the rupture of rock or ice or frozen earth, and many other processes, make their contributions to local and minute movements. For the greater part, these vibrations are superficial in origin, and are soon damped beyond recognition by dispersal and by the inelastic and discontinuous nature of the looser material of the surface. When a temporary rigid crust is formed by freezing, as in winter, these surface vibrations are transmitted with much less loss, and the distances at which the rumble of winter traffic is heard, is a good illustration of the function of continuity and solidity in the conveyance of vibrations.

Earthquakes.

When the tremors spring from sources within the earth itself and are of appreciable violence, they are recognized as earthquakes. The sources of earthquake tremors are various. The most prevalent is probably the fracture of rocks and the slipping of strata on each other in the process of faulting. The interpretation of movements of this class has now been so far perfected that the length and depth of the fault, the amount of the slip, and the direction of the hade are capable of approximate estimation.[1] To the same class belong the movements due to slumping. They are illustrated by the sliding and arrest of great masses of sediment along the steep fronts of deltas, and of the accumulations of deep-sea oozes on steep submarine slopes. Such slumping is, in reality, superficial faulting. Seismic tremors often attend volcanic eruptions, and are then probably attributable to the sudden fracture and displacement of rock by the penetration of lava, or by rapid and unequal heating. They are perhaps also due sometimes to the sudden generation or cooling of steam in underground conduits, crevices, and caverns, the action possibly being in some cases of the "water-hammer" type. In rare instances, probably, the bursting of beds overlying pent-up non-volcanic gases may give origin to earthquakes. A more superficial source of earthquake vibrations is the collapse of the roofs of subterranean caverns.

Seismic vibrations seem to be in part compressional, in part distortional, in part (on the surface) undulatory, and in part irregular. The distortional are especially significant, as they seem to imply a solid medium of transmission.

Points of origin, foci.—It is probable that nearly or quite all earthquake movements start within the upper ten miles of the crust, and most of them within the upper five. Some of the earlier estimates indeed placed the points of origin as deep as 20 or 30 miles, but in these cases the necessary corrections, discussed below, were neglected. Most of the recent and more accurate estimates fall within the limits given.

The method of estimating the depth of the centers of disturbance consists in observing the directions of throw or thrust of bodies at the surface, and in regarding these as representing the lines of emergence of the earthquake-waves. By plotting these lines of emergence, and projecting them backwards to their underground crossings, a first approximation to the location of the focus is reached (the lines *EF′*, Fig. 446). From the nature of the case, the observations of the angles of emergence cannot be very accurate, but an effort is made to limit the error by making the number of observations great.

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[1] Davison. Jour. of Geol., Vol. VIII, p. 301.

Two systematic corrections are to be applied to all such estimates, the one for varying elasticity and density, and the other for varying continuity. Both reduce the estimated depth. In making the correction for varying elasticity, it must be noted that the velocity of vibrations varies directly as the square root of the elasticity, and inversely as the square root of the density. The velocity is also accelerated by increase of temperature. The elasticity, temperature, and density all increase with depth. Theoretically, the increase of velocity due to the increasing elasticity and temperature of increasing depths, overbalances the retardation due to increasing density, and recent observations on the transmission of seismic waves through deep chords of the earth have confirmed this conclusion. The path of the vibration will, therefore, be curved toward the surface, as pointed out by Schmidt and illustrated in Fig. 446, taken from his discussion.[1] From this it is clear

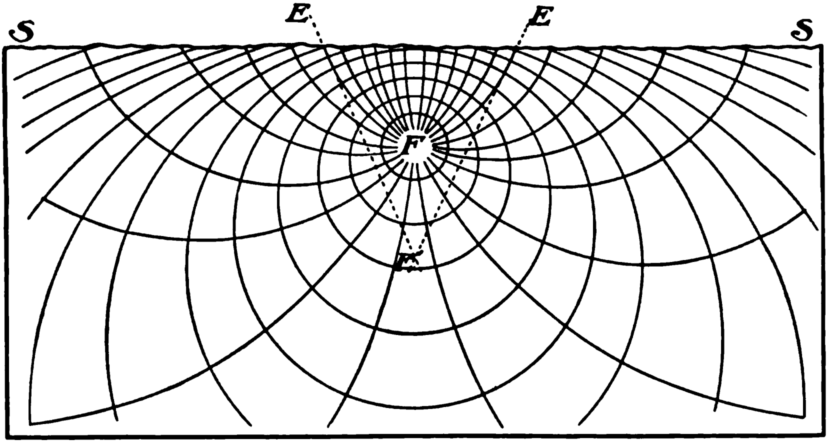

Fig. 446.—Diagram illustrating by closed curves the different rates of propagation of seismic tremors from a focus *F*, and, by lines normal to these, the changing directions of propagation of the wave-front. It will be seen that the paths of propagation curve upwards in approaching the surface. If the lines of emergence, as at *E* and *E*, be projected backwards, as to *F′*, the points of crossing will be below the true focus.

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[1] Nature, October 24, 1895.

that the focus is not so deep as implied by the simple backward projection of the lines of emergence.

A second correction must be made for the differences of continuity of the upper rock in the vertical and horizontal directions. In the outer part of the earth, the continuity in horizontal directions is interrupted by vertical fissures. Were these not usually filled with water, they would soon kill the horizontal component of the seismic wave, and the residual portion would be directed almost vertically to the surface, for the width of the fissures is almost always greater than the amplitude of the seismic vibrations. The water restores the continuity, in a measure, but not perfectly, for the elasticity of water is much less than that of rock. It is clear that in horizontal movement there must be a constant transfer from rock to water and from water to rock, and this must retard, as well as partially destroy, the vibrations. In a vertical direction, however, the rocks rest firmly upon one another, and this gives measurable continuity, the only change being from one layer or kind of rock to another. It seems certain, therefore, that the vertical component of the seismic wave will be less damped and less retarded in transmission than the horizontal. It will, therefore, reach the surface sooner and will have the greater effect on bodies at the surface, not only for the reasons given, but also because it emerges more nearly in the line of least resistance and of freest projection. On this account, a second correction must be added to the correction for elasticity, and this must further reduce appreciably the first estimate of the depth of the focus.

Observation shows that in some way a seismic wave becomes separated in transmission into portions of different natures and speeds, but their interpretation is yet uncertain. These separated portions probably consist of the compressional, the distortional, and the undulatory waves, and perhaps of refractions and reflections of these (see Fig. 448).

A most important recent achievement is the detection and investigation of seismic tremors that appear to have come *through* the earth. The transmission of such waves promises to reveal much relative to the nature of the deep interior, when enough data are gathered to warrant conclusions. The rate of propagation in the central parts is found to be greater than in the outer parts, implying high elasticity within.

The amplitude of the vibrations.—From the very disastrous effects of severe earthquakes, it is natural to infer that the distinctive oscillations must have large amplitude, but in fact it is the suddenness of the vibration, rather than its length, that is effective. Instrumental investigations indicate that the oscillations, after they have left their points of origin, are usually only a fraction of a millimeter in amplitude; at most they seldom exceed a few millimeters. A sudden shock with an amplitude of 5 or 6 millimeters is sufficient to shatter a chimney. It is true that estimates assigning amplitudes of a foot or more have been made, but their correctness is open to serious doubt. It should be understood that it is the length of oscillation of the particles of the sub-surface rock transmitting the vibrations that is referred to, not the movement of the free surface, or of objects on the surface. The throw at and on the surface is much greater. Just as a slight, quick tap of a hammer on a floor is sufficient to make a marble lying on it bound several inches, so a sufficiently sudden rise of the surface of the earth, though but a fraction of an inch, may project loose bodies many feet.

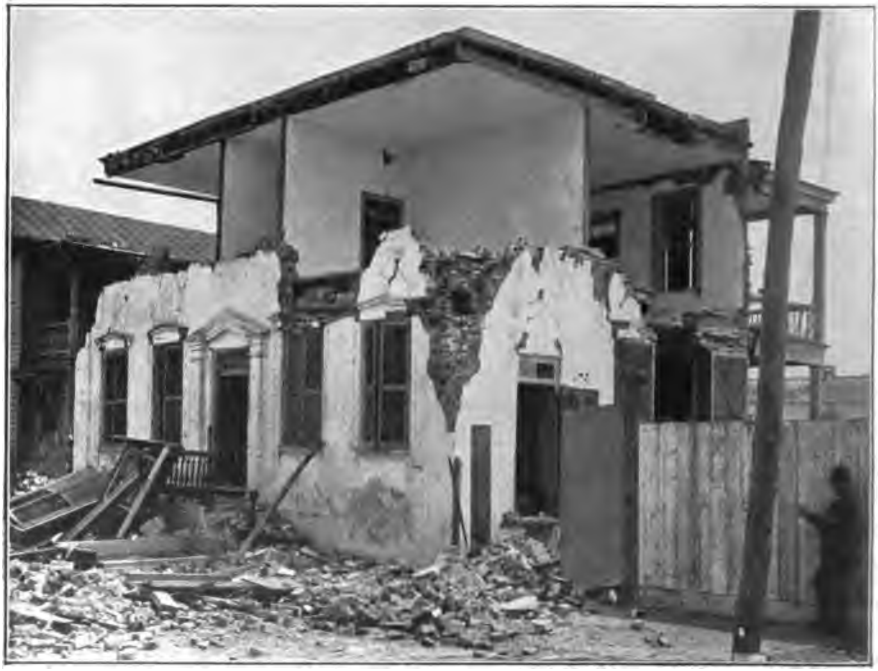

FIG. 447.—Illustration of the destructive effects of the Charleston earthquake, showing definite direction of throw. (W J McGee.)

Destructive effects.—The interpretation of the disastrous results of earthquake shocks has, therefore, its key in the suddenness and strength of rather minute vibrations of the earth-matter, but it is also dependent on the freedom of motion of the bodies affected. The rocks of the deeper zones, where the matter is sensibly continuous, transmit the seismic vibrations without appreciable disruptive effect, so far as known, though the origin of crevices has been assigned to this cause; but bodies at the surface are fractured, overturned, and hurled from their places. The reason is doubtless this: Within a great mass firmly held in place by cohesion and pressure on all sides, the forward motion of a particle develops an equal elastic resistance, and it is quickly thrown back again and the wave passes on. At the surface, where bodies are freer to move, the stroke of the vibration projects the body, and so, instead of vibratory resilience, the chief energy is converted into mass-motion. The tap of a hammer sends an almost imperceptible vibration along the floor, but this vibration may throw a glass ball, beneath which it runs, into the air. So the minute vibrations of earth-matter may travel miles from their origin through continuous substance with little result, and then so suddenly thrust a loose body on the surface, or the base of a column, or the foundation of a house, as to rack it with differential strains, or even to hurl it to destruction. So, too, earth-waves striking the sea-border may thrust the waters off shore by their sudden impact, and the reaction may develop a wave which overwhelms the coast. Such waves may doubtless arise from a sudden stroke of seismic vibrations on the sea-bottom. The great gaping fissures that sometimes open during earthquakes occur oftenest where the surface on one side is less well supported than on the other, as on a slope, or near a bluff-face or a river-channel. When in such situations the earth is once suddenly forced in the direction of least resistance, it is not always met by sufficient elastic resistance to throw it back. Sometimes, however, there is an elastic return, and the fissure closes forcibly an instant after it is opened.

Direction of throw.—Immediately above the point of origin, technically the *epicentrum* or *epi-focal point*, bodies are projected upwards. When crushing takes place in such a case, it is due to the upthrust or to the return downfall. At one side of the epicentrum the thrust is oblique in various degrees, and is usually more destructive, if not too far from the epicentrum. The destructiveness commonly increases for a certain distance from the epi-focal point, and then diminishes. Under ideal conditions, the greatest effects are found where the vibration emerges at an angle of about 45°, but various influences modify this result. Lines drawn through points of equal effect (isoseismals) are not usually regular circles or ellipses about the epicentrum, as they would be under ideal conditions. The various divergencies represent differences of effective elasticity, of surface, and of other influences. As most earthquakes originate from lines, planes, or masses, rather than points, there are doubtless differences of intensity of vibration at different points on the lines, planes, or areas of origin, and these differences introduce inequalities in propagation and in surface effects.

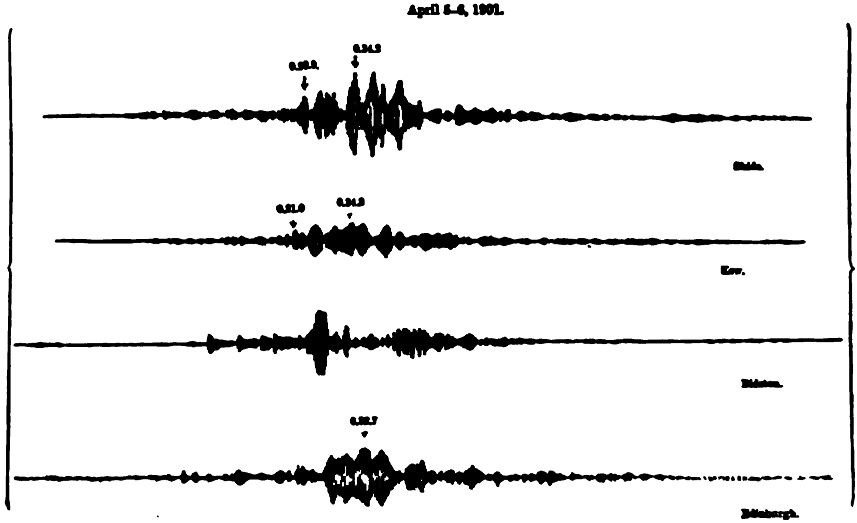

FIG. 448.—Illustrations of the records made by earthquake tremors after distant transmission through the earth. The four diagrams represent the same set of tremors as received at Shide, Kew, Bidston, and Edinburgh in Great Britain. The movement was from left to right. (Milne.)

Rate of propagation.—The progress of a seismic wave varies very greatly. Both experimental tests and natural observations give very discordant results. At present, they justify only the broad statement that the velocity of propagation varies from several hundreds to several thousands of feet per second at the surface. The rate seems to be greater for strong vibrations than for weak ones, and hence it is faster near the origin than farther away. The strength of a vibration dies away, theoretically, according to the inverse square of the distance from the point of origin. Practically there is to be added to this the partial destruction of the vibrations by conversion into other forms of motion.

Sequences of vibrations.—Near the source, the main shocks are apt to come suddenly and to be followed by minor tremors. At a distance there are usually "preliminary" vibrations followed by the main tremors, and these by others of gradually diminishing value. This development is assigned to different rates of propagation, and to refractions and reflections not unlike the prolongation of thunder (see Fig. 448). This deployment of the vibrations is notably developed in the shocks that pass long distances through the earth. The vibrations of the first phase are regarded as compressional, those of the second as distortional, while the largest oscillations which arise still later perhaps come around the surface, and may be undulatory, though their nature is not yet determined.

There is often, however, a true succession of original shocks caused by a succession of slips or ruptures at the source. Sometimes these are exceedingly persistent, running through days, weeks, or even months. In such cases a slow faulting is probably in progress, and little slips and stops follow in close succession. In one instance as many as 600 shocks in ten days have been reported.[1]

Gaseous emanations.—Vapors and gas frequently issue from earthquake rents, and are popularly made to serve as causes, but they are usually merely the earth gases that are permitted to escape by the rending of the ground, or are forced out by readjustment of the shaken beds. Like other subterranean gases, they are often sulphurous, and they are sometimes hot, especially in volcanic regions. Where the shocks are connected with eruptions, the gases may be truly volcanic.

Distribution of earthquakes.—Over large portions of the globe, severe earthquakes are exceedingly rare, but in certain regions they are unfortunately frequent. For the most part, these are volcanic districts, but this is by no means a universal relation. Earthquakes and volcanoes are only in part associates. In general, it may be said that earthquakes are frequent where geologic changes are in rapid progress, as along belts of young mountains, where the stresses are not yet adjusted, or at the mouths of great streams, where deltas are accumulating, or about volcanoes, where temperatures and strains are changing, or on the great slopes, particularly the submarine slopes, where readjustments in response to inequalities of surface stress are in progress. Not a few, however, occur where the special occasion is not at all obvious.

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[1] Milne. The Geog. Jour., Vol. XXI, p. 1. See also Seismology, a more technical work than the same author's Earthquakes.

The Geologic Effects of Earthquakes.

Earthquakes are of much less importance, geologically, than many gentler movements and activities. Disastrous as they sometimes are to human affairs, they leave few distinct and readily identifiable marks which are more than temporary.

Fracturing of rock.—During the passage of notable earthquake waves, the solid rock is probably often fractured (see p. 509), though where it is covered by deep soil the fractures are rarely observable at the surface. Elsewhere the crevices are readily seen, especially if they gape. In a few instances surface-rock has been seen to be thoroughly shattered after the passage of an earthquake, as in the Concepcion earthquake of 1835.[1] Joints which were before closed are often opened during an earthquake. Thus in northern Arizona, not far from Canyon Diablo, there is a crevice traceable for a considerable distance, which is said to have been opened during an earthquake. Locally, it gaps several feet. Other notable earthquake fissures have been recorded in India,[2] Japan, and New Zealand. During an earthquake which shook the South Island of New Zealand in 1848, "a fissure was formed averaging 18 inches in width, and traceable for a distance of 60 miles, parallel to the axis of the adjacent mountain chain."[3] The development of fractures or the opening of joints is sometimes accompanied by faulting. This was the case in Japan during the earthquake of October 28, 1891, when the surface on one side of a fissure, which could be traced for 40 miles, sank 2 to 20 feet. In this case there was also notable horizontal displacement, the east wall of the fissure being thrust locally as much as 13 feet to the north.[4]

Changes of surface.—Circular surface openings or basins are sometimes developed during earthquakes. This was the case during the Charleston earthquake of 1886,[5] and similar effects have been noted elsewhere. These openings often serve as avenues of escape for ground-water, gases, and vapor. They are commonly supposed to be the result of the collapse of caverns, or other subterranean openings, the collapse often causing the forcible ejection of water. Such openings are likely to be formed only where the surface material is incoherent. Sandstone dikes (p. 514) may perhaps be associated in origin with earthquakes.

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[1] Darwin. Journal of Researches, 1845, p. 303.

[2] Oldham. Quar. Jour. Geol. Soc., Vol. XXVIII, p. 257.

[3] Geikie. Text-book of Geology, 4th ed., p. 372.

[4] Kotô. Jour. Coll. Sci., Japan, Vol. V, Pt. IV (1893), pp. 329, 339. Cited by Geikie, loc. cit., p. 373.

[5] An elaborate account of this earthquake is given by Dutton, Ninth Ann. Rept., U. S. Geol. Surv., pp. 209–528.

Earthquakes are likely to dislodge masses of rock in unstable positions, as on slopes or cliffs. They may also occasion slumps and landslides.[1]

Effects on drainage.—The fracturing of the rock may interfere with the movement of ground-water. After new cracks are developed, or old ones opened or closed, the movement of ground-water adapts itself to the new conditions. It follows that springs sometimes cease to flow after an earthquake, while new ones break out where there had been none before. The character of the water of springs is sometimes changed, presumably because it comes from different sources after the earthquake. Joints may be so widened as to intercept rivulets, and the waters thus intercepted may cause the further enlargement of the opening. Illustrations of this sort are furnished by the earthquakes of the Mississippi valley (Lat. 36° to 38°) in 1811–12. Where faults accompany earthquakes, they occasion ponds or falls where they cross streams. Illustrations of both were furnished by the Chedrang River of India after the earthquake of 1897.[2]

Effects on standing water.—Some of the most destructive effects of earthquakes are felt along the borders of the sea. Thus the great sea-wave of the Lisbon earthquake (1755) and that of the earthquake which affected the coasts of Ecuador and Peru in 1868 are examples. Such waves have been known to advance on the land as walls of water 60 feet in height. They are most destructive along low coasts, for here the water may sweep much more extensively over the land. The great loss of life during an earthquake has usually been the direct result of the great waves. Lakes are also affected by earthquakes, their waters sometimes rising and falling for several hours after the initial disturbance, but lake-waves are much feebler than those of the sea, and are not often destructive.

Earthquake shocks are sometimes remarkably destructive to the life of lakes and seas. Thus during the Indian earthquake of 1897, "fishes were killed in myriads as by the explosion of a dynamite cartridge . . . and for days after the earthquake, the river (Sumesari) was choked with thousands of dead fish . . . and two floating carcasses of Gangetic dolphins were seen which had been killed by the shock."[1] This wholesale destruction of life is of interest, since the surfaces of layers of rock, often of great age, are sometimes covered with fossils of fish or other animal forms, so numerous and so preserved as to indicate that the animals were killed suddenly and in great numbers, and their bodies quickly buried. It has been suggested that such rock surfaces may be memorials of ancient earthquake shocks.[2]

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[1] Cross. Twenty-first Ann. Rept., U. S. Geol. Surv., Pt. II, Chap. V.

[2] Oldham. Report on the Indian Earthquake of June 12, 1897, p. 138. Mem. Geol. Surv. of India. Cited by Geikie, loc. cit., p. 374.

Changes of level.—Permanent changes of level sometimes accompany an earthquake. Thus after the earthquake of 1822 "the coast of Chili for a long distance was said to have risen 3 or 4 feet."[3] Similar results have occurred on the same coast at other times, and on other coasts at various times. Depression of the surface is perhaps even more common than elevation. Thus on the coast of India all except the higher parts of an area 60 square miles in extent were sunk below the sea during an earthquake in 1762. Widespread depression in the vicinity of the Mississippi in Missouri, Arkansas, Kentucky, and Tennessee accompanied the earthquakes of 1811 and 1812. Some of the depressed areas were converted into marshes, while others became the sites of permanent lakes. Reelfoot Lake, mainly in Tennessee, is an example. Change of level is involved wherever there is faulting, and faulting is probably rather common in connection with earthquakes.

Changes of level are not confined to the land. Where earthquake disturbances affect the sea-bottom in regions of telegraph cables, the cables are often broken. In such cases notable changes have sometimes been discovered and recorded when the cables were repaired. Striking examples are furnished by the region about Greece.[4] In one instance (1873) the repairing vessel found about 2000 feet of water where about 1400 feet existed when the cable was laid. In another instance (1878) the bottom was "so irregular and uneven for a distance of about two miles, that a detour was made and the cable lengthened by five or six miles." In still another case (1885) the repairing vessel found a "difference of 1500 feet between the bow and stern soundings." These records point to sea-bottom faulting on a large scale.

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[1] Oldham, loc. cit., p. 80.

[2] Geikie. Text-book of Geology, 4th ed., p. 375.

[3] Ibid., p. 376.

[4] Forster, Seismology, 1877. Summarized in the Am. Geol., Vol. III, 1889, p. 182.

It is probably no nearer the truth to say that changes of level result from earthquakes than to say that earthquakes result from changes of level. The two classes of phenomena are probably to be referred to a common cause.[1]

SLOW MASSIVE MOVEMENTS.

It is a far cry from the intense and inconceivably rapid oscillations of the earthquake, to the excessively slow subsidences of continents, or even the slow wrinkling of mountain folds. Not infrequently rivers wear down their channels across a mountain range as fast as it rises athwart them. The movements of continents are even more deliberate. But, far apart as these contrasted movements are, in rate and method, they are associated in ultimate causation, and the earthquake shock is often merely an incident in the formation of a mountain range or in the subsidence of a continent.

The great movements are usually classed (1) as continent-making (epeirogenic) and (2) mountain-making (orogenic). They may also be classed as (1) vertical movements and (2) horizontal movements, and dynamically, as (1) thrust movements and (2) stretching movements. It is to be understood that these distinctions are little more than analytical conveniences, for continental movements are often at the same time mountain-making movements; vertical movements are usually involved in horizontal movements, and stretching usually takes part in the processes in which thrust predominates, and *vice versa*. But where one phase greatly preponderates, it may conveniently give name to the whole.

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[1] The literature of seismology is very extensive. Some of the more general treatises are the following: Mallet, Brit. Assoc., 1847, Part II, p. 30; 1850, p. 1; 1851, p. 272; 1852, p. 1; 1858, p. 1; 1861, p. 201; and The Great Neapolitan Earthquake of 1857, 2 Vols., 1862; A. Perrey, Mém. Couronn. Bruxelles, XVIII (1844), Comptes Rendus, LII, p. 146; R. Falb, Grundzüge einer Theorie der Erdbeben und Vulkanenausbrüche, Graz, 1871, and Gedanken und Studien über den Vulkanismus, etc., 1874; Pfaff, Allgemeine Geologie als exacte Wissenschaft, Leipzig, 1873, p. 224; Schmidt, Studien über Erdbeben, 2d ed., 1879, and Studien über Vulkane und Erdbeben, 1881; Dieffenbach, Neues Jahrb., 1872, p. 155; M. S. di Rossi, La Meteorologia Endogena, 2 Vols., 1879 and 1882; J. Milne, Earthquakes and other Earth-movements (contains a bibliography), 4th ed., 1898; Seismology, ibid., 1898; Dutton, Earthquakes, 1904.

Records of earthquakes have been preserved more or less fully in several countries, especially in recent years. A few of the more accessible publications where these

Present movements.—Critical observations on seacoasts show that some shores are slowly rising and some slowly sinking *relative to the ocean-level.* We do not certainly know what their movements are relative to the center of the earth; very possibly all may be sinking, one set faster than the other, the ocean-surface also going down at an intermediate rate. Theoretically, all might possibly be rising, one set faster than the other, the ocean also rising at an intermediate rate, though this is extremely improbable. One set may be actually rising relative to the center of the earth, and the other sinking, while the ocean-level is stationary, or nearly so. This is the way in which we are accustomed to interpret them. A general shrinkage of the earth, however, is probably going on, carrying down land-surface and sea-surface. It has been urged by Suess[1] that the general shrinkage is so great that the local upward warpings and foldings never equal it, and that the real movements are all downward, though in different degrees. This is probably the general fact at least. Over against this is the popular disposition to regard earth movements generally as "upheavals." There is also a predilection for regarding the rigid land as moving and the mobile sea-level as fixed. In reality, the sea is an extremely adaptive body that settles into the irregular hollows of the lithosphere, and is shifted about with every warping of the latter. Whatever change affects the capacity of its depressions affects also the sea-level. If they are increased, the sea settles more deeply into them; if they are decreased, the sea spreads out more widely on the borders of the land. The one thing that gives a measure of stability to the sea-level is the fact that all the great basins are connected, and so an *average* is maintained. A warping down in one part of the sea-bottom may be offset by an up-

records are found are cited below: California earthquakes, Perrine, Bull. 147, U. S. Geol. Surv.; Earthquakes of the Pacific Coast, Holden, Smithson. Misc. Coll., No. 1087, 1898; Records of recent earthquake movements in Great Britain since 1890 are published by Davison in Quar. Jour. Geol. Soc., Geol. Mag., and Nature; Records of earlier earthquakes are found in the reports of the Brit. Assoc. (Mallet), in the Edinburgh New Philos. Jour., Vols. XXXI–XXXVI (Milne), and in Trans. of the Roy. Irish Acad., 1884 and 1886 (O'Reilly); The Earthquakes of Scandinavia have been recorded in volumes of the Geol. Fören, Förhandl.; Records of other continental European earthquakes are found in Gerland's Beiträge zur Geophysik, 1895, 1900, and 1901; Neues Jahrb., 1865–71; Zeitschr. Naturwissen. (1884), (Credner); Bericht. k. Sachs. Geol. Wissen., 1889 and 1900 (Credner); Jahrb. Geol. Reichsanst., 1895 and 1897; Tschermak's Min. Mitth., 1873, and later; Transactions of the Seismological Soc. of Japan. An index to these Transactions is given at the end of Milne's Seismology.

[1] Antlitz der Erde. Vol. 1, p. 136.

ward warping somewhere else in the 72% of the earth's surface covered by the ocean, and so it is only the *sum total of all changes* in the sea-bottoms and borders that effects the common level. Thus it happens that, notwithstanding its instability and its complete subordination to the lithosphere, the sea-level is the most convenient basis of reference, and has become the accepted datum-plane. If there were some available mode of measuring the distance of points from the center of the earth, it would give absolute data and absolute terms, and would reveal much that is now uncertain respecting the real movements of the surface. For convenience, however, since absolute terms are impracticable, the ordinary language of geology, which represents movements as upward and downward, according to their relations to the sea-level or to the average surface, will be employed. Notwithstanding this concession to convenience in the use of terms, it is of the greatest importance to form, and to constantly retain, true fundamental views.

Fundamental conceptions.—The existence of any land at all is dependent on the inequalities of the surface and of the density of the lithosphere, for if it were perfectly spheroidal and equidense, the hydrosphere would cover it completely to a depth of about two miles. Not only are inequalities necessary to the existence of land, but these inequalities must be *renewed from time to time*, or the land area would soon, geologically speaking, be covered by the sea. The renewal has been made again and again in geological history by movements that have increased the inequalities in the surface of the lithosphere. With each such movement, apparently, the oceans have withdrawn more completely within the basins, and the continents have stood forth more broadly and relatively higher, until again worn down. This renewal of inequalities appears to have been, in its great features, a *periodic movement*, recurring at long intervals. In the intervening times, the sea has crept out over the lower parts of the continents, moving on steadily and slowly toward their complete submersion, which would inevitably have been attained if no interruption had checked and reversed the process. These are the great movements of the earth, and in them lies, we believe, the soul of geologic history and the basis for its grand divisions. The reasons for this will appear as the history is followed, and its most potential agencies are seen unfolding themselves. At the same time, there have been numerous minor surface movements in almost constant progress. While these two classes of movements have been associated, and are perhaps due in the main to the same causes, they are sufficiently different in some of their dynamic aspects to be separated in treatment.

Nearly Constant Small Movements.

Innumerable gentle warpings have affected nearly every portion of the surface of the globe at nearly all stages of its history. Not only during the periods of great movements were there countless minor and gentler movements, but at times of relative quiescence there were slow swellings and saggings of the surface of the lithosphere. They sometimes affected small areas and sometimes large ones, and they were sometimes of upward phase and sometimes of downward. They were the immediate agencies in locating and controlling the deposition of stratified rocks, though they rest back on the great movements for their working conditions. Very slow sinkings of sea-borders have permitted deposition to go on in shallow water for long periods without being interrupted by the local filling of the sea. Very slow swellings of land tracts, relative or absolute, have permitted erosion to supply material for such sedimentation for long periods without exhausting the sources. Very slow upward warpings in one region and downward warpings in another have shifted the borders of the land and sea, and with them the areas of erosion and deposition. Thus have arisen overlaps and unconformities of strata and diversities in their distribution from stage to stage. Such movements may have amounted to a few inches, or a few feet, or a few fathoms per century. Downward movements have sometimes affected a considerable section of a continent, letting in a shallow *epicontinental* sea upon it, such, for example, as the North Sea upon the northwestern border of the continent of Europe, and Hudson Bay upon the northeastern part of North America. Similar movements seem to have extended the seas even more widely upon the surface of the land in times past, as attested by the great transgressions of the ocean-borders and the great epicontinental spread of strata. Notwithstanding their great breadths, the epicontinental seas were generally shallow. Similar gentle warpings of upward phase rescued the bottoms of shallow seas from submersion, and inaugurated erosion; or they bowed base-leveled lands upward, and rejuvenated their streams and inaugurated a new cycle of denudation. Often they connected continents previously separated by shallow straits, and thus inaugurated inter-continental migrations of land life, while they stopped inter-oceanic migration.

The gentleness and frequency of these movements is attested by the character of the sediments and by their relations to one another, as will be seen in the study of the sedimentary series.

Reciprocal features.—These minor warpings show a notable tendency to be reciprocal. If one area is bowed up, another near by is bowed down. If the continents settle, the oceans rise on their borders. If the land is cut down, the sea is filled up. There is an important phase of this deserving especial note. Certain tracts have been slowly bowed upwards into long land swells, the streams being rejuvenated and degradation hastened. Adjacent tracts have been slowly bowed downwards into long parallel troughs which received the wash from the adjacent swells, and thus became tracts of exceptional sedimentation. Such a tract of parallel swell and sag, if our interpretation be correct, developed along the Atlantic border of North America in the Paleozoic era. By the slow upward warping of the swells, the feeding-grounds of the streams were maintained, and the sags were filled about as fast as they sank. Thus a great depth of sediment was laid down in the course of an era measured by millions of years. So in other regions, especially near the borders of the continents, there have been similar reciprocal movements, giving at once feeding-grounds for the streams and lodgment-grounds for the sediments, side by side in parallel belts. It is a common view that these belts of deep sedimentation were the forerunners of mountain formation, and that they determined the formation of the mountains. In view of the grounds for doubting the efficiency of so superficial an agency in mountain formation, which will appear as we go on, it may be well to hold this view in abeyance, and to dwell on the reciprocal nature of the action, in which the upward bowing that gave the feeding-grounds is as vital a factor as the sagging that accommodated the sedimentation. It is important to recognize that in so far as the crust was weak enough to yield to these gentler forces, it was not strong enough to accumulate the great stresses necessary to form mountain ranges, and further, that in so far as the stresses were eased by the gentle warping, they could not be accumulated for the later work of mountain-folding. It is nevertheless probable that the conditions which located the gentle swelling and sagging also located the mountain-folding.

The Great Periodic Movements.

Mountain-forming movements.—Along certain tracts, usually near the borders of the continents, and at certain times, usually separated by long intervals, the crust was folded into gigantic wrinkles, and these constitute the chief type of mountains, though not the only type. The characteristic force in this folding was lateral thrust. The strata were not only arched, but often closely folded, and sometimes intensely crumpled. In extreme cases, like the Alps, the folds flared out above, giving overturn dips and reversed strata, as illustrated in the chapter on Structural Geology, pp. 501–511. In these cases there was an upward as well as a horizontal movement, for the folds themselves were lifted; but the horizontal thrust so much preponderated, and was so much the more

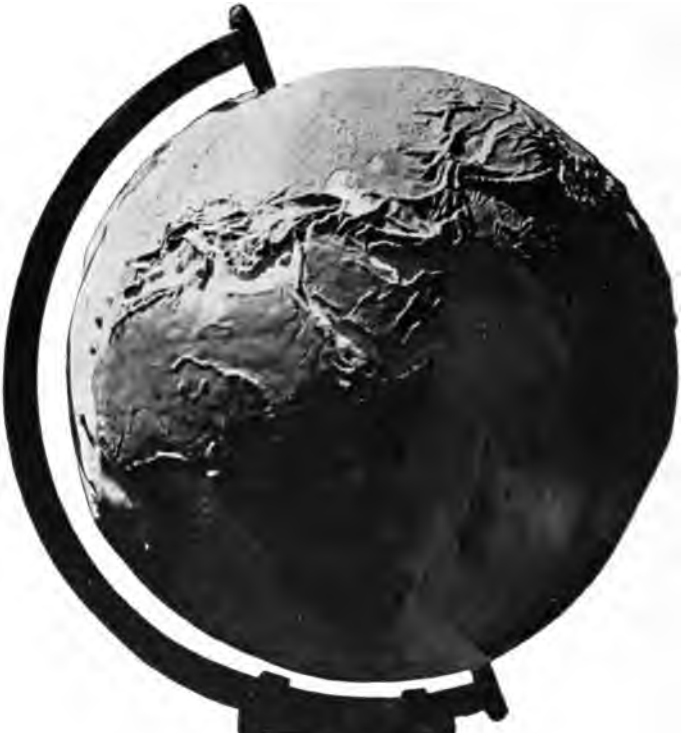

FIG. 449.—The great Eurasian mountain tract. Jones Relief Globe. (Photo. by R. T. Chamberlin.)

remarkable, that the upward movement was overshadowed. It is well to note, however, that these mountain ranges are crumpled *outward* and not *inward*, as might be expected if they resulted simply from the shrinkage of the under side of a thin shell. The folds are sometimes nearly upright and symmetrical, and sometimes inclined and asymmetrical, as illustrated in the chapter referred to. Where the folds lean, the inference has been drawn that the active thrust came from the side of the gentler slope, the folds being pushed over toward the resisting side, and this seems to be commonly true. The original attitude of the beds, however, has much to do with the character of the folds.[1] By a slight change in the mode of thrust, sheets of paper may be so pushed as to lean forward or backward at pleasure. The leaning of the folds seems, therefore, a doubtful criterion for determining the direction of the active movement. Mountains of the thrust type usually consist of a series of folds nearly parallel to each other, the whole forming an anticlinorium.

Distribution of folded ranges.—The prevailing location of this class of mountains is so generally near the borders of the continents that the relation is probably significant. Dana[2] long ago called attention to the fact that the greatest mountain ranges stand opposite the greatest ocean-basins, and he connected the elevation of the one with the depression of the other. One of the most notable exceptions to this relation is the complex system of southern Europe, from the Pyrenees to the Caucasus, and another is the Altai and connected ranges (Fig. 449). The Urals and not a few minor ranges are also exceptions. It is probably better to regard the crumpled tracts as lying on the borders of great segments of the earth that acted essentially as units, and to regard the relationship to the sea as a coincidence that is only in part causal.[3]

Plateau-forming movements.—Another leading phase of crustal movement is the settling or rising of great blocks of the crust, as though by vertical rather than horizontal force. The western plateau of North America and the great plateau of Thibet are gigantic examples. The American plateau embraces numerous blocks which, while they have been elevated together, are individually tilted in their own fashion. At the surface, they are separated by fault-planes, but below, some of them, and perhaps most of them, pass into flexures. Most of these flexures are of the monoclinal type (p. 516), which dynamically means much the same as a fault; but some of them may be of the compressive type, without inconsistency with vertical fault-relief above. Research has not yet covered thoroughly any great plateau, and knowledge of this class of movements is less complete than that of folding by lateral thrust, and it has a less ample place in the literature of the subject. The plateau-forming movements are, however, much more massive than the mountain-folding movements, and stand next in magnitude to the continent-forming movements. Plateaus may be regarded as smaller platforms superposed on the continental platforms.

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[1] Eugene A. Smith. Underthrust Folds and Faults, Am. Jour. Sci., Vol. XLV, 1893, pp. 305–6.

[2] Manual of Geology, 3d ed., p. 23.

[3] For discussions of folds, see Van Hise, Sixteenth Ann. Rept. U. S. Geol. Surv., Pt. I, pp. 603–632; and Willis, Thirteenth Ann. Rept., Pt. II, pp. 217–296.

In the ocean-basins, there appear to be raised platforms of the plateau type, and there are remarkable "deeps" that have the aspect of anti-plateaus.

Continent-forming movements.—True continent-forming movements appear to have antedated the earliest known sediments. As far back as we can read the sedimentary record, the continents seem to have been well established, and there is little evidence that they have since been fundamentally changed. It is true that some very eminent geologists have rather freely connected formations on one continent with formations having similar faunas on an opposite continent, by a hypothetical conversion of the intervening ocean-bottoms into land or shallow water; but most such faunal relations can be explained almost equally well by migration around the coasts, or at most by mere ridge-connections. The paucity, if not total absence, of abysmal deposits in the strata of the continents, taken with the persistence of terrestrial and coastal faunas, leaves little room for assigning an interchange of position between abysmal depths and continental elevations, and *vice versa.* Dynamic considerations also offer grave difficulties. The doctrine of the persistence of continents probably ought not to be pushed so far as to exclude shallow water, or even land, connections between South America, Antarctica, Australia, India, and South Africa, directly or indirectly, at certain stages of geological history. Without forming final conclusions as to the measure of the change which the continents have suffered during known geological history, it is safe to conclude that the continents and ocean-basins were in the main formed very early in the earth's history, and that subsequent changes have consisted chiefly in the further sinking of the basins and the further protrusion of the land, save as the latter has been cut down by erosion. Incidentally, the ocean-basins have probably been extended and the continents restricted. On the other hand, the continents have been built out on their borders by wash from the land, and the waters of the ocean have been somewhat lifted by the deposition of sediment in their basins. It is estimated that the cutting away of the present continents, and the deposition of the material in the ocean-basins, would raise the sea-level about 650 feet. (R. D. George.)

Relations of these movements in time.—The folding movements seem to have had extraordinary prevalence in the earliest ages, for the Archean rocks are almost universally crumpled, and often in the most intricate fashion. There is no sign that the folding was then limited to the borders of the continents; it seems rather to have affected the whole continental surface. After the beginning of the well-known sedimentary series, crumpling appears to have taken place chiefly at long intervals, thus marking off great time-divisions, and to have been confined at any given stage to certain tracts, chiefly on the borders of great segments of the earth's crust.

Concerning the plateau-forming movements in the past, knowledge is very meager, as the detection of plateaus of ancient times is more difficult than the detection of folds. Gentle warpings have apparently been in progress at all times.

Relations of vertical to horizontal movements.—The downward movements are unquestionably the primary ones, and the horizontal ones are secondary and incidental. The fundamental feature is doubtless central condensation actuated by gravity, and the master movements are the sinkings of the ocean-basins. The great periodic movements that made mountains and plateaus, and changed the capacity of the ocean-basins, probably started with the sinking of part or all of the ocean-bottoms. In the greater periodic movements, probably all the basins participated more or less, but some seem to have been more active than others. For example, in the last great mountain-making period, the Pacific basin seems to have been more active than the Atlantic, while in the similar great event at the close of the Paleozoic, the opposite seems to have been true. The squeezing up of the continents doubtless took place simultaneously with the settling of the basins. The true conception is perhaps that the ocean-basins and continental platforms are but the surface forms of great segments of the lithosphere, all of which crowd toward the center, the stronger and heavier segments taking precedence and squeezing the weaker and lighter ones between them. The area of the more depressed or master segments is almost exactly twice that of the protruding or squeezed ones. This estimate includes in the latter about 10,000,000 square miles now covered with shallow water. The volume of the hydrosphere is a little too great for the true basins, and it runs over, covering the borders of the continents. The amount of the overflow fluctuates from time to time, and may be neglected in a study of the movements and deformations of the lithosphere.

The squeezed segments.—The great protruding segments show a tendency toward rude triangularity. They are (1) the Eurasian, now strongly ridged on the south and east, and relatively flat on the northwest; (2) the African, rather strongly ridged on the east, but less abruptly elevated on the west and north; (3) the North American, now strongly ridged on the west, more gently on the east, and relatively flat at the north and in the interior; (4) the South American, strongly ridged on the west and somewhat on the northeast and southeast.

The foregoing form the major group. The minor group embraces (5) the Antarctic segment, not as yet sufficiently known to be well defined, and (6) the Australian, broadly reniform rather than triangular. To these are perhaps to be added (7) the largely submerged platform that stretches from Sumatra and Java on the southwest to the Philippines on the northeast, and is attached to India on the northwest; and (8) Greenland, which, though closely associated with North America, is partially separated by a rather deep depression.

The depressed or master segments.—The great sunken segments show a tendency to assume roughly polygonal, rather than triangular, forms. This accords with the primary place assigned them, since, in a spherical surface divided into larger and smaller segments, the major parts should be polygonal while the minor residual segments are more likely to be triangular. The major segments are (1) the Pacific, (2) the Indian, (3) the North Atlantic, and (4) the South Atlantic. These form the principal group, while (5) the Arctic deeps (not including the shallow epicontinental portions), (6) the Mediterranean, (7) the Caribbean, and (8) the chain of deep pits between the Philippine ridge and the Bornean platform, constitute a subordinate group.

Each member of the minor group is an irregular chain of depressed pits rather than a single continuous deep, unless the Arctic depression, of which little is now known, proves an exception. They lie between the greater segments at what may be conceived to be points of *critical working relations*, and are accompanied by small elevated blocks. The Caribbean, the Mediterranean, and the Bornean regions are the seats of the greatest present volcanic and related activities.

In a general view, there are then four great sunken quadrilaterals and four great elevated triangles, with minor attendants in each class. Lest fondness for simplicity and symmetry lead too far, we must hasten to observe that the *dimensions* are not alike in either class. The Pacific segment is more than twice the size of any other basin segment, and four times that of the North Atlantic. The Eurasian triangle is more than twice the average size of the other land segments, and nearly three times that of the South American. Nor is there any large common divisor of approximate accuracy. This is not at all strange if the earth be regarded as a body of somewhat heterogeneous composition which naturally shrank in rather irregular segments. On the other hand, this irregularity is somewhat strange if the earth has evolved from a very homogeneous and symmetrical, primitive, fluid state. It is also a serious consideration in any theory that appeals to crystalline form, or analogy, as in the doctrine of a tetrahedral earth.

Roughly approximated in millions of square miles, the major depressed segments are as follows: the Pacific, 60, the Indian, 27, the South Atlantic, 24, and the North Atlantic, 14, leaving 8 for minor depressions. The elevated segments are Eurasian, 24, African, 12, North American, 10, and South American, 9, leaving 10 for the minor blocks.

If these segments be regarded as the great integers of body-movement, two-thirds of them taking precedence in sinking and the other third in suffering distortion, it is easy to pass to the conception of sub-segments, moving somewhat differently from the main segments, so as to aid in their adjustment to one another, and thus to the conception of plateaus and deeps. It is easy also to pass to the conception of mutual crowding and crumpling at the edges of these segments, accompanied by fracture and slipping. These conceptions perhaps represent the true relations between the massive movements of the abysmal and continental segments, as well as the less massive plateau-forming movements and the mountain-forming distortions. The mountains and plateaus are probably the incidental results of the great abysmal and continental readjustments.

The great movements are probably to be attributed to stresses that gradually accumulated until they overcame the rigidity of the thick massive segments involved, and forced a readjustment. In accumulating these stresses, some local yielding on weak lines and at special points was an inevitable incident in distributing more equably the accumulating stresses. So, also, the first great readjustments probably left many local strains and unequal stresses which gradually eased themselves by warpings, minor faultings, etc., so that some minor movements were a natural sequence of the great movements. But there were doubtless many local and superficial causes, such as irregular gains and losses of heat, regional loading and unloading, solution, hydration, etc., that have caused local or regional movement, and which have little to do with the great deformations of the earth's body. As implied above, the gentle, nearly constant movements probably fall mainly into a different category from the great periodic movements. Both will be considered further.

The differential extent of the movements.—Between the highest elevation of the land and the lowest depth of the ocean, there is a vertical range of nearly twelve miles. There may have been higher elevations, relatively, in past times, but probably not deeper depressions; and so, if we assume that the surface was once perfectly spheroidal, this may be taken as a maximum expression of *differential* movement, not *absolute* vertical movement. From the Thibetan plateau, where a considerable area exceeds three miles in height, to the Tuscarora deep, where a notable tract exceeds five miles in depth, the range is eight miles, which may fairly represent the vertical range of rather massive differential movement. From the average height of the continents to the average abysmal bottoms of the oceans the range is nearly three miles, which may be taken as the differential movement of the great segments. Under certain hypotheses of the origin and early history of the earth, to be sketched later, the surface is not assumed to have been perfectly spheroidal originally, and hence the present irregularities do not necessarily imply so great differential movement.

If the protruding portions of the lithosphere were graded down and the basins graded up to a common level, this level would lie about 9000 feet below the ocean-surface. This equated level is the best basis of reference for relative segmental movements. Referred to this datum plane, the continents, having an area about half as great as that of the ocean depths, have been squeezed up relatively about two miles, and the basins have sunk about one mile from the ideal common plane. The total downward movement, representing the total shrinkage of the earth, is quite unknown from observation. It is probably very much greater than the differential movement, as will appear from theoretical considerations as we go on.

The extent of the *lateral* movements has a peculiar interest, for it bears theoretically on the shrinkage of the earth. Every mile of descent of the crust represents 6 miles (6.28) shortening of the circumference. If the vertical movements were limited to the *relative* ones just named, the mile of basin descent would give but little more than 6 miles of surplus circumference for lateral thrust and crumpling. How far does this go in explaining the known facts? By measuring the folds of the Alps, Heim has estimated the shortening represented by them to be 74 miles.[1] Claypole estimated the shortening for the Appalachians in Pennsylvania, not including the crystalline belt on the east, at 46 miles;[2] McConnel placed that of the Laramide range in British America at 25 miles,[3] and LeConte that of the Coast range in California at 9 to 12 miles.[4] These estimates must be corrected for the thickening and thinning of the beds in the process of folding, for the composite character of the folds, and for the effects of shearing and faulting. These will in part tend to increase and in part to decrease the estimates. The first effect of horizontal thrust is to close up all crevices and compact the beds as much as they will stand without bending. A part of the unusual thickness which the beds of folded regions commonly show is probably due to this edgewise compression. In experiments on artificial strata made to illustrate foldings (Fig. 449*a*), the thickening of the layers is a very appreciable part of the process, though probably natural beds do not thicken in equal proportion. After the beds have been closely folded and the thrust is athwart them, they are thinned and stretched on the imbs of the fold. How far this and other causes of extension offset nitial compression is undetermined, and is differently estimated. It seems highly probable from the nature of the case that the edgewise compression which resulted from sustaining the full stress before the beds bent, was much greater than the crosswise compression on the limbs of the folds, which came into action only after the stress had been largely satisfied by folding.

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[1] Mechanismus der Gebirgsbildung, p. 213.
[2] Am. Nat., Vol. XIX, p. 257, 1885.
[3] Geol. Surv. of Canada, p. 33 D, 1886.
[4] Elements of Geology, 5th ed., p. 266.

Whatever the correction, and whatever the probable errors of the above estimates, the amount of shortening involved in folding is large.

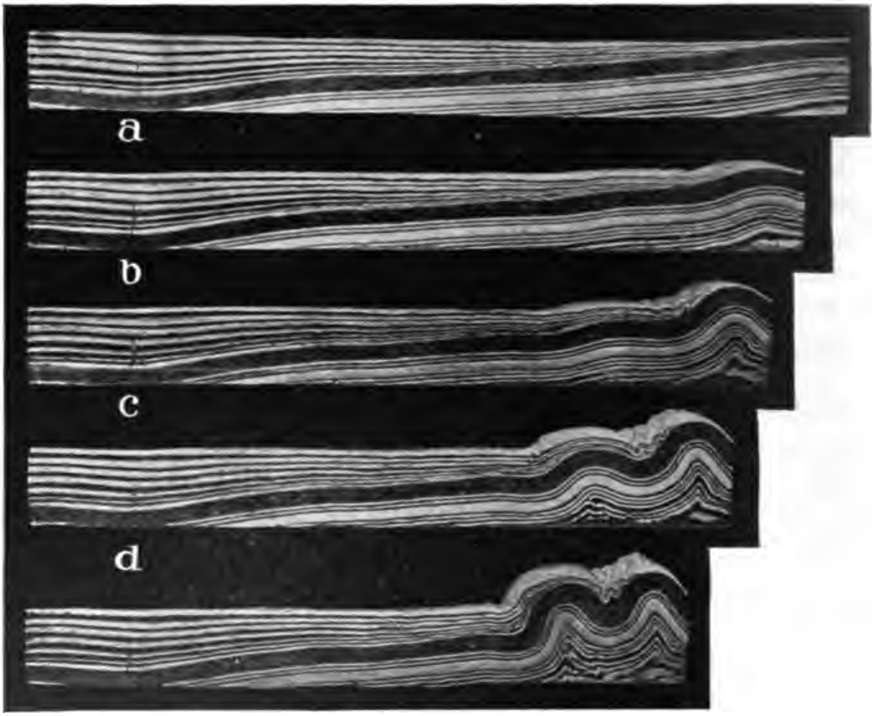

FIG. 449a—Illustrations of Willis' experiments in the artificial representation of mountain folding. The sections were formed of layers of wax of different colors, and were mechanically compressed from the right. The upper section shows the original state, and the offsets of the succeeding sections at the right indicate the amount of shortening. (Thirteenth Ann. Rep. U. S. Geol. Surv.)

The estimates given are merely those for certain periods of folding, and represent only that portion of the compression of the circumference which was concentrated in a given mountain range. The whole shortening of a circumference is to be found by adding together all the transverse foldings on a given great circle, following it about the globe at right angles to a given folded tract. In so doing, it will be seen that the belt does not usually cross more than one or two strongly folded tracts of the same age, from which it is inferred that the shortening on each great circle was largely concentrated in a few tracts running at large angles to each other, to accommodate the shrinkage of the globe in all directions. If the folding in a main range crossing any great circle is doubled, it will probably represent roughly the shortening for that entire circle for that age. If one is disposed to minimize the amount of folding, the estimate may perhaps be put roundly at 50 miles, on an entire circumference, for each of the great mountain-making periods. If, on the other hand, one is disposed to give the estimates a generous figure so as to put explanations to the severest test, he may perhaps fairly place the shortening at 100 miles, or even more. For the whole shortening since Cambrian times, perhaps twice these amounts might suffice, for while there have been several mountain-making periods, only three are perhaps entitled to be put in the first order, that at the close of the Paleozoic, that at the close of the Mesozoic, and that in the late Tertiary. The shortening in the Proterozoic period was considerable, but is imperfectly known. The Archean rocks suffered great compression in their own times, and probably shared in that of all later periods, and if their shortening could be estimated closely, it might be taken as covering the whole. Assuming the circumferential shortening to have been 50 miles during a given great mountain-folding period, the appropriate radial shrinkage is 8 miles. For the more generous estimate of 100 miles, it is 16 miles. If these estimates be doubled for the whole of the Paleozoic and later eras, the radial shortening becomes 16 and 32 miles, respectively.

THE CAUSES OF MOVEMENT.

General Considerations.

The volume of the earth is at all times dependent on two sets of antagonistic forces, (1) the attractive or centripetal, consisting of gravity and the molecular and sub-molecular attractions, and (2) the resistant forces—which are not necessarily centrifugal—consisting of heat and the resistant molecular and sub-molecular forces.

1. *The centripetal agencies.*

Gravity.—The most obvious of the concentrating forces is gravity, and in most questions relating to great segmental movements, it has been thought sufficient to consider gravity alone, but it is by no means certain that this does not lead to serious error. In studying the causes and effects of earth movements, it is necessary to consider both gravitational *energy* and gravitational *force*. Gravitational *energy* is greatest when the mass is most widely dispersed, and least when most concentrated. Gravitational *force* is greatest when the mass is most concentrated, and least when most dispersed. The gravitational energy of the earth matter was at its maximum when it was most widely diffused in the supposed nebulous condition. It will perhaps reach its minimum at some future period when the shrinkage shall reach its limit. In passing from an expanded condition to a more concentrated condition, potential energy, or energy of position, is transformed into other forms of energy, chiefly heat. The heat thus developed is an important factor in the earth's dynamics. The total amount of gravitational energy involved in the earth's evolution is unknown, for neither the maximum dispersion of the earliest state, nor the ultimate condensation, is known. It is not difficult, however, to compute the amount of transformation of gravitational energy into heat, or other forms of energy, during a given degree of condensation. If a mass equal to that of the earth were originally infinitely scattered, the gravitational energy given up by it in condensing into a homogeneous sphere of the earth's present size would, if all transformed into heat, suffice to raise the temperature of an equal mass of water 8900° C. (Hoskins), or an equal mass of rock (specific heat of .2), 44,500° C. If the mass were more condensed toward the center, as is the actual case, the heat would be considerably greater. If the condensation toward the center followed the Laplacian law (p. 564), the heat would be sufficient to raise the earth mass 48,900° C., assuming its specific heat to be .2, which is about the average specific heat of rock at the surface (Lunn). A further shrinkage of one mile would transform an additional amount of gravitational energy into heat about equal in amount to Tait's estimate of the loss of heat from the surface of the earth in 100,000,000 years (see p. 572). If the radial shrinkage has been 32 miles, or even 16 miles, the amount of heat generated is very much greater than the estimated loss from the surface.

How much gravitational energy can possibly be transformed into heat and other forms of energy in the future, can only be computed by making assumptions as to the possible extent of further contraction, and that involves hypotheses as to the atomic and sub-atomic constitution of the earth's matter, and its behavior under the prodigious pressures of the earth's interior. All shrinkage develops added gravitational force and further tendency to shrinkage, which follows when the heat generated by the shrinkage is lost; and where the process may end, in a body of the dimensions of the earth, is beyond present determination. If there were no limit to the density that might be attained, it would be impossible to assign any limit to the energy that might be transformed. It has usually been assumed that contraction could not go on indefinitely because the atoms would come into actual contact, and prevent further increase of density. This conception rests on the recently prevalent hypothesis of the atomic constitution of matter; but the more recent hypotheses that substitute multitudes of revolving corpuscles or electrons for irreducible atoms, do not carry the same presumption of a rigorous limit to condensation. It is not therefore prudent to try to set such a limit, or to make it a feature in the dynamical doctrines of the earth. It is even less prudent to try to measure the limit of future conversion of the gravitational energy of the sun into heat, and so to set a limit to the habitability of the earth.

The *force* of gravity may be defined as the effort of gravitational energy to change into other forms of energy. It is most familiarly expressed in terms of weight, which is the resultant of the gravitational force of the whole earth upon a given portion. Weight is determined by the distances and directions of the given portion from all parts of the attracting mass, the amount of the attraction being directly as the mass and inversely as the square of the distance, modified by the direction. It is greatest about 610 miles below the surface, where it is 1.0392 times that at the surface. Below this point it declines, and at the center it is zero. The sum total of the earth's gravitative force at the present time is equivalent to about 6×10^{21} tons. This gives rise to a pressure of about 3,000,000 atmospheres at the center of the earth.

Gravitational force is also expressed in terms of the earth's ability to accelerate the velocity of falling bodies at its surface, which is now approximately 32 feet per second. For certain purposes, the force of gravity may be better pictured by means of the velocity required to overbalance it, which is 6.9 miles per second; e.g. a body shot away from the surface at a speed exceeding 6.9 miles per second, would escape from the control of the earth if the influence of the atmosphere and other bodies is neglected; while a body shot away at less than this speed would return to the earth.

Molecular and sub-molecular attractions.—In addition to gravity, there are at least three additional classes of attractive agencies whose laws appear to differ from those of gravity, viz. cohesion, chemical affinity, and sub-atomic attraction, using these terms in their comprehensive generic senses. The thought has been entertained that these might be reducible to forms of gravity in ulterior analysis, but it does not appear from existing evidence that the laws of their attractions are conformable to the Newtonian law of the inverse square of the distance, to which gravity conforms. Apparently the forces of the molecular, atomic, and sub-atomic attractions increase at higher rates, and have individual peculiarities of action quite different from gravity. It would be of the utmost service to geological philosophy if these laws of molecular and sub-molecular attractions were firmly established, and could be applied to the conditions of heat and pressure under which the matter of the interior of the earth exists. In the absence of such determinations, we can do little more than recognize that the matter of the interior of the earth tends to condense itself by the aid of molecular and sub-molecular attractions, supplemental to the attraction of gravity.

Cohesion and crystallization.—The force of gravity between small bodies is exceedingly feeble, but it is cumulative, every particle in a mass attracting every other particle, so that in great masses the force becomes enormous. In cohesion, and probably in the other molecular and sub-molecular attractions, the particles attract very strongly the particles with which they are in close relations, but beyond minute distances their effects are insensible. The force of crystallization is felt for a very short distance from the crystal, and "mass action" is probably dependent on a function of similar kind, acting at a very small distance, but the range of these forces is very limited in comparison with that of gravity.

Rock matter, as a rule, tends to become crystalline by the assembling of like molecules in systematic order. The general effect is condensation, though this is not universally the case, for in some instances the crystalline arrangement results in expansion. The crystallizing force may be regarded as a specialized variety of cohesion which usually cooperates with gravity to produce increased density. In cases of expansion it seems clear that the organizing force does not act according to the law of gravity. The intensity of the force exhibited in the formation of ice illustrates the superiority of the molecular force over the gravitative force in small masses; but in a planet of ice of very moderate dimensions, the internal pressure of gravity would overcome the crystalline force, which illustrates the superiority of gravity in large masses.

While the crystalline force may thus in exceptional cases operate against gravity, it is known that in most cases it not only operates with it, but is controlled by it, in this sense—that where a substance has two forms of crystallization, it will take the denser one when the pressure is great. The inference is that if the less dense form of crystallization takes place under slight pressure, and subsequently the pressure is greatly increased, the form of crystallization will change from the less to the more dense.[1] It is probable that in general those forms of molecular arrangement will be assumed in the deep zones which give the greatest density, and this probably includes concretionary, colloidal, and other forms of aggregation, as well as crystallization.

Diffusion.—The same law probably holds relative to diffusion, though in a molecular sense diffusion is the opposite of crystallization, for in crystallization, like comes to like, while in diffusion the molecules distribute themselves among those of unlike nature. Diffusive action, quite familiar in gases and liquids, takes place to some extent in solids. The molecules of plates of gold and lead brought into intimate contact under pressure mutually diffuse among one another. So gases seem to be very generally diffused or "occluded" in rocks, though the nature of this relation is imperfectly determined. It is known that pressure upon gases promotes their diffusion through liquids and solids. It is inferred that pressure upon a solid tends to the diffusion of the entrapped gases within it, but it is not to be inferred from this that pressure upon rock promotes the absorption of gases into it, but rather the opposite. It is probable that great pressure with high heat promotes the diffusion of entrapped gases or other diffusible substances through the rock-mass, and at the same time tends to their extrusion along lines of least resistance; but this is an inference rather than a demonstration.

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[1] Van Hise. Bull. Geol. Soc. of Am., 1897, Vol. IX, p. 291.

Chemical combination.—The general effect of chemical combination under pressure is greater density. In reversible reactions capable of conditions of chemical or physical equilibrium, pressure invariably favors the formation of the denser of any possible products.

Sub-atomic forces.—Recent investigation has made it probable that atoms are composite, embracing many exceedingly minute bodies—corpuscles or electrons—in a state of extremely high activity and possessed of marvelous energy notwithstanding their minuteness. This discovery possesses deep interest to the geologist because it seems to reveal sources of energy of almost incalculable potency, some portions of which at least are being constantly freed and added to the previously recognized supplies of energy. Attempts have been made during the past few decades to limit the habitable age of the earth, both retrospectively and prospectively, by the smallness of the sum total of energy derivable from gravity. In these estimates slight recognition has been given to the resources of molecular and atomic energy, and none at all to the possibilities of sub-atomic energy. It would be going quite too far to assume that these sub-atomic energies are all available for the perpetuation of habitable conditions on the earth or in the solar system, but we are doubtless justified in appealing to them as an offset to all dicta restricting the period of the earth's habitability by supposed insufficiencies of energy deduced merely from the estimated resources of gravity. The banishment of the idea of the atom as a minute, incompressible, undecomposable sphere takes away the theoretical limit of compressibility, and by so doing cuts away the groundwork for assigning definite limits even to the resources of gravity, since, as already indicated, unlimited condensation gives theoretically unlimited transformation of the potential energy of gravity.

While we must await with such patience as we can command the development of fuller knowledge concerning the nature and laws of the molecular, atomic, and sub-atomic energies, and their applicability to the activities resident in the interior of the earth, it is permissible even now to assume that, besides the simple compressive action of gravity, there are at work varied forms of molecular aggregation, of atomic combination, and perhaps of sub-atomic change, tending toward increased density, and that the ulterior limit of these processes is quite undetermined. The condensational forces are now restrained at certain temporary limits by the antagonistic resistant forces, some of which, such as heat, are the products of the condensational forces, and are gradually being dissipated, permitting further condensation. Where the process may ultimately end, we dare not attempt to say. On the other hand, we are not compelled to accept assigned limits that seem to be inconsistent with the phenomena which the earth actually presents.

2. *The resisting agencies.*

Heat.—The most familiar of the active agencies that resist condensation is heat. Upon this the existing volume of the earth is immediately dependent, in some large part at least. As this heat is dissipated, the earth shrinks. This shrinkage increases the force of gravity, and hence the internal pressure increases, and, if further compression takes place as the result of this increased pressure, additional heat is developed, which checks further condensation until it is dissipated. It is this kind of creative and self-checking action that determines the volume of great gaseous bodies like the sun. Though their matter is far from its ultimate density, and their self-gravity is enormous, they condense slowly, because, with every stage of condensation, heat is generated which antagonizes gravity and checks condensation, until at least a part of the heat is radiated away. As the force of gravity increases with every stage of condensation, the heat developed to hold it in check must increase, and hence the famous law of Lane, that a gaseous body like the sun grows hotter as it condenses. This law holds good while the body remains in a gaseous state in which the maintenance of the volume is essentially dependent on heat. When a body becomes liquid or solid, its volume is dependent in part on forms of resistance other than heat, and the force of the law is abated, though the principle still holds good. In small solids, the principle has little application, since the force of self-gravity is slight compared to the resisting forces, and very little new heat is generated as the body loses that which it has; but in large bodies, like the earth, where the condensational forces are enormous and the internal temperature is very high, it is not improbable that the heat generated at every stage of condensation is relatively large. It has been inferred by some students of the phenomena that the conditions in the interior of the earth are essentially those of gaseous matter, so far as molecular relations are concerned, because the temperatures are thought to be above the critical temperatures of the substances composing it. If this be true, the new heat generated with each stage of condensation is large. However this may be, it seems safe to infer that in so far as the volume of the interior mass is dependent on heat resistance, *the loss of existing heat leads to the generation of new heat.* The amount of this new heat must be enough, together with the residual heat and the other forces of resistance, to match the new condensational forces. The molecular and sub-molecular forces of resistance other than heat, are probably responsible for some large part of the resistance to the increased condensational force, but how much is not determined.

All resistance perhaps due to motion.—As now interpreted, the force of resistance of heat is due to the impact of the flying particles of the heated matter. The other forms of resistance to compression have not usually been interpreted in this way, but the tendency of recent investigation is to place them in the same dynamic class. A cold solid body offers resistance to compression that is in no obvious way dependent on heat motion. In small bodies this resistance is immeasurably greater than the self-gravity of the body. It is so great that it can only be partially overcome by any force which human ingenuity can bring to bear upon it. This form of resistance has thus, not unnaturally, come to be regarded as approximately immeasurable, and perhaps as grading into actual immeasurability, and as resting back upon the actual contact of irreducible atoms. But the recent researches which have developed grounds for the conception that even the atoms are composite, lead to the further conception that their resistance to compression is dependent on the movement of their constituent corpuscles or electrons. This encourages the broad conception that the whole of the resistance to compression arises from molecular, atomic, and sub-atomic *motions*, of which heat is merely one form.

While all this is yet on the frontier of physical progress, these conceptions may well be recognized in framing interpretations of the agencies which determine the volume of the earth, and which control the changes that take place in it from age to age. The result of their combined action at any stage is *a state of temporary equilibrium* between gravity, aided by the molecular, atomic, and sub-atomic attractions, on the one hand, and heat, aided by the molecular, atomic, and subatomic resistances, on the other. The vital problem is to ascertain *the original condition of balance* between these antagonistic forces, and the changes which have affected that balance since. The original state of balance is necessarily a matter of hypothesis, and the best that can be done at present is to picture as clearly as possible the different hypotheses that have been entertained, and the different consequences that logically flow from them. The most important factor in the case is the original amount and distribution of internal heat.

ALTERNATIVE VIEWS OF ORIGINAL HEAT DISTRIBUTION.

The hypothetical modes of origin of the earth will be treated in the historical section. Suffice it here to say that one view is that the earth was once gaseous, passed thence into a liquid, and later into a solid state. Under this view, there are two hypotheses as to the original distribution of internal heat, dependent on the mode of solidification. According to the one, solidification began at the surface after convection had brought the temperature of the whole mass down nearly to the point of congelation; according to the other, solidification began at the center at a high temperature, because of pressure, and proceeded thence outwards. The former only has been much developed in the literature of the subject, though the latter is now generally regarded as the more probable.

Another view of the globe's origin is that the earth was built up gradually by the infall of matter, bit by bit, at such a rate that though each little mass became hot as a result of its fall, it cooled off before others fell on the same spot, the rain of matter not being fast enough to heat up the whole mass to the melting-point. Under this view, the internal heat arose chiefly from compression due to the earth's gravity.

A clear conception of the three hypotheses of thermal distribution which rest on these two views of the origin of the earth is important to the further discussion.

1. **Thermal distribution on the convection hypothesis.** — It was formerly the prevailing opinion that the molten condition of the earth persisted in the interior until after the crust had formed, and that solidification proceeded from the surface downwards. It was a natural corollary of this view that, previous to the beginning of solidification, convection stirred the liquid mass from center to circumference and equalized the temperature so that the whole mass cooled down equably until it approached the point of solidification and became too viscous for ready convection. The temperature should, therefore, have been nearly the same from center to surface at the stage just preceding incipient solidification. This conception forms the basis of most discus-

sions involving internal temperatures.[1] The famous studies of Lord Kelvin are based on the assumption of a uniform initial temperature of 7000° Fahr.[2] Other temperatures have been assumed in similar studies by others, but the results do not differ materially. On this hypothesis there would be no deep-seated change of temperature until a temperature-gradient, extending to the deeper horizons, had been developed by surface cooling. In the earliest eras, the loss of heat would be felt solely in the outer zone. By surface cooling, a temperature gradient would be slowly developed, and gradually changed from age to age, as shown by the curved lines in Fig. 450, each of which shows the temperature at the successive stages stated in the legend. The computations for these curves were based on the methods and assumptions of Lord Kelvin. The two lower curves represent greater periods than those usually assigned by geologists to the whole history of the earth. It will be seen that the modification of the original temperature line extends only about 160 miles below the surface for the 100,000,000-year period, only about 240 miles for the 237,000,000-year period, and only about 320 miles for the excessive period of 600,000,000 years. The superficial nature of the whole thermal problem under this hypothesis is thus made clear and impressive.

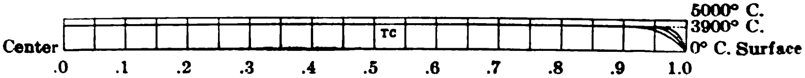

FIG. 450.—Diagram showing the original distribution of heat assumed by the convection hypothesis and the modifications of this distribution near the surface in successive long periods The base-line of the figure represents divisions of the earth-radius with center at the left and surface at the right. The vertical lines represent temperatures ranging from 0° C. to 5000° C. The assumed initial temperature 3900° C. (7000° F.) is represented by the horizontal line TC, full at the left and dotted at the right to indicate the original extension of the initial temperature to the surface. The upper curve at the right shows how much the temperature will have been modified at the end of 100,000,000 years, computed according to the method of Lord Kelvin. The middle curve shows the change at the end of 237,000,000 years, and the lower curve the change at the end of 600,000,000 years. Similar curves may be found in an article by Clarence King, Am. Jour. of Sci., XLV, 1893, p. 16.

After the outer shell had cooled so as to be in approximate equilibrium with the environment of the earth, it suffered practically no contraction.

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[1] See Woodward's address, Mathematical Theories of the Earth, Proc. Am. Assc. for Adv. Sci., 1889, pp. 59–63.

[2] Nat. Phil. Thompson and Tait, Pt. II, p. 477. See also Popular Lectures and Addresses, 1894, II, p. 313.

So also it appears from the diagram that there was practically no contraction below 160 miles up to the end of the 100,000,000-year period, because cooling had not yet reached that depth. Between these two non-contracting horizons the greatest rate of contraction at the close of the 100,000,000-year period lay about 60 miles below the surface. The contraction of this middle zone, while the outermost shell and the interior body remained constant, is held to have developed a state of horizontal thrust in the outer shell, because this shell, being too large for the shrinking subcrust, tended to settle, and to crowd upon itself horizontally. The wrinkling and other modes of deformation of the outer part of the earth are referred, under this view, to the thrust so developed. This is the view which has been most generally accepted.

Level of no stress.—As the outer shell is thus held to be in a state of thrust while the zone below is in a state of shrinkage, there must be, between these two zones, a *level of no stress*, where there is neither compression nor stretching. Above this level, the thrust increases to the surface, and below it, the stretching increases to the depth of most rapid change of temperature, below which it decreases and finally vanishes at the lower limit of temperature change. In the earliest stages of cooling, the level of no stress must have been near the surface, and must have descended gradually as the cooling proceeded. The depth of this level has been repeatedly computed on the basis of assumed times and rates of cooling. Fisher, assuming the temperature of solidification to have been 4000° Fahr. and the period of cooling 33,000,000 years, computed its depth at only $\frac{1}{10}$ of a mile below the surface.[1] T. Mellard Reade, with somewhat different assumptions, placed it at 2 miles after 100,000,000 years of cooling.[2] Davison (1897) placed it at 2.17 miles,[3] and G. H. Darwin at 2 miles after the same period.[3] In a later computation, based on the assumption that the coefficient of dilatation increases with the temperature, Davison placed the level of no stress at 7.79 miles, and stated that if the coefficient of conductivity and the initial heat also increased downwards, the zone would lie still deeper. To suppose the initial heat to increase downwards, however, is to abandon the hypothesis we are now considering. These computations seem to show that, at the very utmost, the level of no stress, under this hypothesis, lies at a very slight depth, and that the thrust zone above is, therefore, very shallow. This should be kept constantly in mind in all deductions drawn from this hypothesis. If the thickness of the thrust zone be taken at 8 or 10 miles, it will apparently be conceding to the view all that can legitimately be claimed for it.

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[1] Physics of the Earth's Crust, Fisher, p. 95.

[2] Origin of Mountain Ranges, T. Mellard Reade, p. 125.

[3] Phil. Trans. Roy. Soc., Vol. 178, pp. 231–49.

2. **Thermal distribution on the hypothesis of central solidification.**—When the previous conception was first formed, the effect of pressure on the melting-points of lavas was neglected, as little or nothing was known on the subject. Experiment, however, has shown that pressure, as a rule, raises the melting-points of lavas, and out of this has grown the doctrine that the earth solidified first at the center, where the pressure was greatest, and gradually congealed outwards. Barus has shown that the melting-point of diabase,[1] selected as a representative rock, rises directly with the pressure. If this rate holds good to the center of the earth, the melting temperature of diabase there would be 76,000° C. (136,800° F.). The range of the experiment is, however, very small compared with the range of the application, and little confidence can be felt in the special numerical result reached. The rate of rise of the fusion-point may be much changed as the extraordinary conditions of the deep interior are invaded. Still there is good ground for the hypothesis that solidification took place at some very high temperature at the center, because of the very great pressure there. The inference then is that when the temperature of the center of the supposed molten globe reached the appropriate point, solidification began there, and that it took place at lesser depths in succession as the appropriate temperatures were reached. This view *excludes convection* in the successive zones from the center outward after the time when their temperatures of solidification were reached, or after these were approached sufficiently near to develop prohibitive viscosity. Some loss of heat from these horizons would be suffered while the outer parts were solidifying, but on account of the exceedingly slow conductivity of rock, it is improbable that the amount of loss would be sufficient to change the general character of the internal distribution of heat previous to solidification at the surface, the time when the existing phase of the earth's history by hypothesis began. Fig. 451 shows the theoretical distribution of heat under this view. The consequences of this assumption are very important to geological theory and, carried out to their logical conse-

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[1] Amer. Jour. Sci., 1893, 3d series, Vol. 45, p. 7.

quences, lead to the conclusion that cooling and shrinkage *affected the deep interior of the earth,* for the high central heat must have been constantly passing out toward the surface. Instead, therefore, of the contraction being concentrated in and limited to the outer 200 miles or so, as under the preceding hypothesis, it was deeply distributed. The contraction within the outer zone would be less than under the preceding view, because the flow of heat from within would partially offset the flow outwards, and a corresponding part of the contraction would be distributed below.

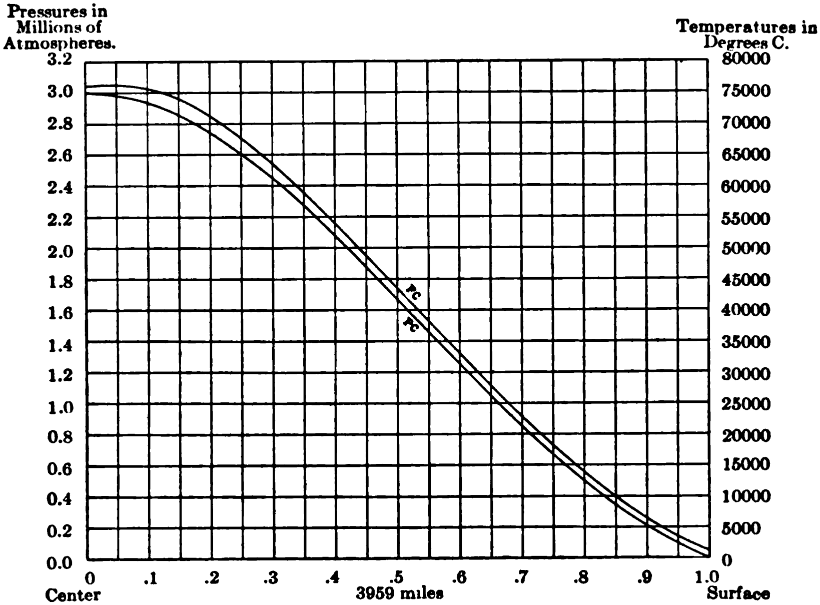

FIG. 451.—Diagram illustrating the internal temperatures of the earth when it first became solid, under the hypothesis that it solidified from the center outward, and assuming that the fusing-point rose directly as the pressure, in accordance with Barus' experiments with diabase. The divisions of the base-line represent fractions of the earth's radius. The divisions of the vertical lines represent pressures in atmospheres at the left, and temperatures in degrees C. at the right. The lower curve, PC, represents the interior pressures, ranging from one atmosphere at the surface to 3,000,000 atmospheres at the center, derived from Laplace's law of density. The upper curve, FC, represents the fusion-points of diabase at the various depths and pressures, and hence the temperatures at which the interior would become solid at the various depths, or, in other words, the initial temperatures of the solid earth. The lower curve is derived from Slichter; the upper is formed by directly plotting the temperatures given by Barus (Am. Jour. Sci., 1893, p. 7).

3. Thermal distribution under the accretion hypothesis.—The accretion hypothesis assumes that the internal heat was gradually developed from the center outwards as the earth grew and the internal compression was progressively developed. The heat, therefore, continued to rise at the center as long as compression continued, or at least as long as the compression was sufficient to generate heat faster than it was conducted outwards. As the conduction of heat through rock is exceedingly slow, the central heat may be assumed to have continued to rise so long as the infall of matter caused appreciable compression. In the same way, heat was generated progressively in the less central parts, and these parts also received the heat that passed out from beneath. It is assumed under this hypothesis that the degree of interior compression stands in close relation to interior density, for while there would probably be some segregation of heavier matter toward the center and of lighter toward the surface by means of volcanic action and internal rearrangement under stress differences, the interior density is regarded as due mainly to compression. The distribution of internal pressure and density generally accepted is that of Laplace, who assumed that the

COMPUTED PRESSURES, DENSITIES, AND TEMPERATURES WITHIN THE EARTH BASED ON LAPLACE'S LAW.

Distance from center in terms of radius.	Pressure in megadynes per sq. cm.[1]	Density.	Temperature in degrees C.
1.00	0	2.80	0
.95	97,000	3.37	320
.90	215,000	3.95	1,110
.85	353,000	4.54	2,190
.80	510,000	5.13	3,470
.75	684,000	5.71	4,880
.70	874,000	6.28	6,350
.65	1,077,000	6.84	7,860
.60	1,289,000	7.38	9,360
.55	1,507,000	7.90	10,830
.50	1,727,000	8.39	12,250
.45	1,944,000	8.84	13,590
.40	2,154,000	9.26	14,840
.35	2,353,000	9.64	15,980
.30	2,535,000	9.98	17,000
.25	2,698,000	10.27	17,880
.20	2,836,000	10.51	18,610
.15	2,947,000	10.70	19,190
.10	3,029,000	10.84	19,610
.05	3,078,000	10.92	19,870
.00	3,095,000	10.95	19,950

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[1] Essentially the same as atmospheres.

increase of the density varies as the square root of the increase of the pressure. This law gives a distribution of density that accords fairly well with the phenomena of precession of the equinoxes, which require that the higher densities of the interior shall be distributed in certain proportions between the center and the equatorial protuberance whose attraction by the sun and moon causes precession. The increases in pressure, density, and temperature have been computed as follows by Mr. A. C. Lunn,[1] the average specific gravity of the earth being taken at 5.6, the surface specific gravity at 2.8, and the specific heat at .2.

The temperatures are shown graphically in Fig. 452, in which the curves of pressure and density are also given. The nature of the curve of temperature is such that, if the thermometric conductivity of the material is uniform at all depths, the temperature *will fall in the deeper portions and rise in the outer ones*, excluding the surface portions subject to outside cooling. The curve indicates that the rising temperature would affect somewhat more than 800 miles of the outer part of the spheroid, or about half its volume, i.e. the inner half during the initial period ,had a falling temperature and the outer half, except the immediate surface, a rising temperature. This introduces a very singular feature into the problem, for the outer zone must shrink to fit the inner portion that is losing heat, while its own material is expanding because of its increase of temperature. A double distortional effect must result.[2] If the conductivity of the dense interior is greater than that of the outer parts, the effect is intensified. The redistribution of heat resulting from this unequal flowage would in time change the curve so that more nearly equal flowage would result. It would probably take a very long period for this to be effected, on account of the very slow conductivity of rock.

The accretion hypothesis assumes that, during the growth of the earth, large amounts of heat were carried by volcanic action from deeper horizons to higher ones and to the surface, and that this still continues at a diminished rate. It assumes that whenever the interior heat raised any constituent of the interior matter above its fusing-point under the local pressure, it passed into the liquid state, *and was forced outwards by the stress differences to which it was subjected,* unless its specific gravity was sufficiently high to counterbalance them. It is conceived that the more fusible portions were liquefied first, and that in so doing they

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[1] The pressures and densities here given are essentially the same as those previously worked out by others and already published. The temperatures are the results of recent preliminary computations made under the auspices of the Carnegie Institution, and are subject to change on further study. They are based on the assumption that the increase in density is due to compression. They are in general accord with the results previously reached by Dr. F. R. Moulton (see "A Group of Hypotheses Bearing on Climatic Changes," by T. C. Chamberlin, Jour. of Geol., 1897, p. 674). The Rev. O. Fisher, in the Am. Jour. of Sci., 1901, p. 420, gives much higher results.

[2] Attention was called to this feature by Chamberlin in a paper before the Geol. Soc. of Am. at Rochester, December, 1901.

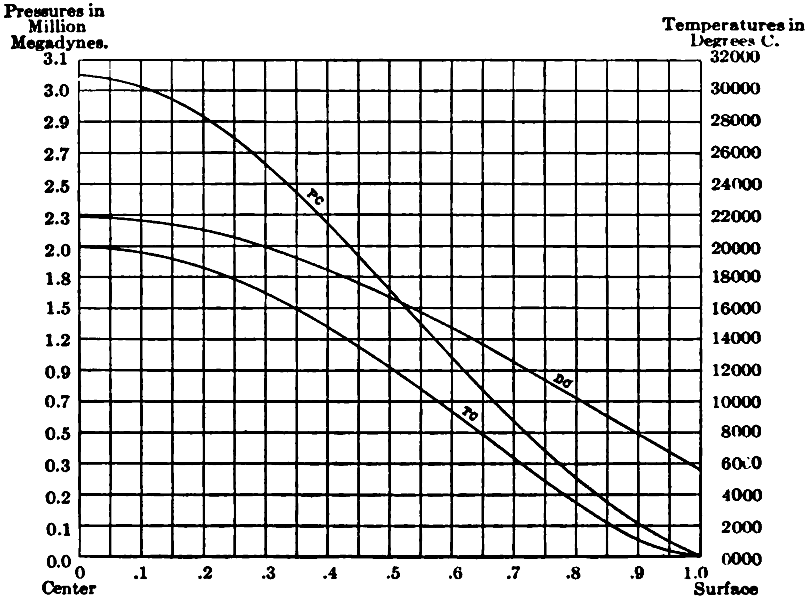

FIG. 452.—Diagram illustrating the distribution of temperature under the accretion hypothesis (neglecting the heat from infall and other external sources). The divisions of the base-line represent fractions of the earth's radius. The vertical divisions represent both pressure in megadynes per sq. cm., nearly the same as atmospheres per sq. in., at the left, and temperatures in degrees C. at the right. It is to be noted that the temperature scale is 2000° C. per division, while that of Fig. 451 is 5000° C. per division. The upper curve at the left, PC, is the pressure curve. The middle curve, DC, is the density curve, beginning at 2.8 at the surface and reaching nearly 11 at the center. The lower curve, TC, is the temperature curve, rising from the surface temperature, 0° C., at the right, to 20,000° C. at the center. It is to be noted that the portion of this curve at the left representing the deeper part of the earth is convex upwards, while the portion at the right is concave. It will be seen that the gradient increases from the center to a point between .6 and .7 radius, and then decreases, and that between .8 radius and the surface, a distance of about 800 miles, the decrease is notable. This means that with an equal coefficient of conductivity the flow from the center outward to .6 or .7 radius will be faster than the flow from .8 radius to the surface, neglecting the immediate surface effects of external cooling. These curves were worked out by Mr. Lunn.

absorbed the necessary heat of liquefaction and began to work their way outward, carrying their heat into higher horizons and temporarily checking the development of more intense stresses in the lower horizons. They thus served to keep the temperature there below the fusion-point of the remaining more refractory substances. Meanwhile the extruded portions were raising the temperatures of the higher horizons into which they were intruded or through which they were forced to pass. There was thus, it is thought, an automatic action that tended to reduce the heat-curve to the fusion-curve. *The actual curve of internal temperature may, therefore, be practically the fusion-curve.* This is identical with the curve supposed to arise from solidification by pressure from the center outward under the molten hypothesis, except so far as the two would vary as the result of variations in the distribution of matter, which would not be quite the same under the two hypotheses. The curve of fusion deduced by an extension of the results of Barus' experiment has been given. It is necessary to recognize that the rate of rise of the fusion-point may, and very likely does, change in the deep interior. The curve given represents much higher temperatures in the central parts than those given by Lunn's computations from compression, which seem inherently more probable than the higher ones.

As astronomical and seismic evidences strongly favor the view that the earth is rigid throughout, they lend support to the view that the interior retains its rigidity by the extrusion of liquid matter practically as fast as it is formed, and that this progressive extrusion adjusts the temperature to that which is consistent with solidity.

The bearing of this conception becomes evident on consideration. The shrinkage of the earth from loss of heat by conduction and by the extrusion of molten rock, affects the deep interior as well as the more superficial zones. *It is even possible that the shrinkage may originate chiefly in the deeper zones.* The postulated transfer of fluid rock from the deeper parts to the more superficial ones lessens the heat in the former, and adds to that in the latter. The postulated greater flow of heat from the deeper half to the outer half, than from the latter outward, gives a concordant result. If the conductivity of the deeper and denser material is appreciably greater than that of the more superficial and less dense material, as seems probable, this effect is intensified. The distribution of compressibility at the existing state of condensation may possibly be such that more new heat is generated by shrinkage in the outer parts than in the inner. Neither of these conceptions can be affirmed as actually taking place. They merely lie within the range of reasonable hypothesis in the present state of experimental data. What the real truth is must be left to further research. Present effort may be regarded as temporarily successful if it forms consistent conceptions of the applicable hypotheses, and of their consequences.

Recombination of material.—One other peculiarity of the accretion hypothesis must be recalled here. The incoming bodies must probably be assumed to have fallen in promiscuous order, and hence to have been indiscriminately mingled in the growing earth. As they became buried deeper and deeper and their temperatures and pressures were raised, much recombination, chemical and physical, may be presumed to have followed. As already noted, these changes would probably give increased density in the main. The material being, however, in a solid state, the rearrangement would be slow and its persistence in time indeterminate, and it may yet be far from complete. It is not improbable, therefore, under this hypothesis, that some notable part of the recent shrinkage of the earth has been due to the continued rearrangement of its heterogeneous internal matter. This would not be equally so in an earth derived from a molten mass, for the required adjustments of the material should have taken place while in the fluid state before solidification.

Comparison of the hypotheses.—By comparing the three hypotheses of the early states of the earth's temperature, it will be seen that there is a radical difference, thermally, between the first and the last two. The first assumes a nearly uniform distribution of internal temperature, and hence, owing to the exceedingly slow rate of conduction, limits the movements and deformations of the crust, so far as dependent on heat, to very superficial horizons. The second and third views agree in postulating changes of temperature in the deep portions, as well as in the superficial, and hence involve the central portion of the earth in the great movements and deformations. It is not to be supposed that this of itself necessarily increases the sum-total of the effects of contraction, for, given a certain loss of heat from the surface, it may be relatively immaterial whether this loss arose from a large reduction of temperature in a shallow zone, or a small reduction of temperature in a deep zone, for, except as the coefficient of expansion varies, the total shrinkage would be the same. But *the difference in distribution* makes a radical difference in the resulting movements, for, in the first case, the movements are in a weak superficial shell that cannot accumulate great stresses, and hence must yield practically as fast as the stresses arise, while, in the second case, the stress-accumulating power of the thick segments may be great, and the stresses may gather for long periods and give rise to great cumulative results at long intervals. In this respect the last two views have much in common, though they differ in other important particulars.

With this general background of hypothesis, we may now turn to the direct evidences of the distribution of internal temperature which observations near the surface afford. Unfortunately they are limited to a mere film, as it were, little more than $\frac{1}{4000}$ of the radius of the earth.

OBSERVED TEMPERATURES IN EXCAVATIONS.

As the earth is penetrated below the zone of seasonal changes by wells, mines, tunnels, and other excavations, the temperature is almost invariably found to rise. The rate of rise, however, is far from uniform. If we set aside as exceptional the unusually rapid rise near volcanoes and in other localities of obvious igneous influence, the highest rates are still six times the lowest. A large number of records have been collated by the Committee on Underground Temperatures, of the British Association for the Advancement of Science. These range from 1° F. in less than 20 feet to 1° F. in 130 feet, with an average of 1° F. in 50 to 60 feet, which has usually been taken as representative. The more recent deep borings that have been carefully measured with due regard to sources of error indicate a slower rate of rise. Some of the more notable records are as follows:[1]

	Depth.		Rate of rise.	
Sperenberg bore (Germany)	3492	feet.	1° F. in 51.5	feet.
Schladeback bore (Germany)	5630	"	1° F. in 67.1	"
Cremorne bore (N. S. Wales)	2929	"	1° F. in 80	"
Paruschowitz bore (Upper Silesia)	6408	"	1° F. in 62.2	"
Wheeling well (W. Va.)	4462	"	1° F. in 74.1	"
St. Gothard tunnel (Italy–Switzerland)	5578	"	1° F. in 82	"
Mt. Cenis tunnel (France–Italy)	5280	"	1° F. in 79	"
Tamarack mine (N. Mich.)	4450	"	1° F. in 100	"[1]
Calumet and Hecla mine (N. Mich.)	4939	"	1° F. in 103	"[1]
Ditto, between 3324 feet and 4837 feet			1° F. in 93.4	"

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[1] These are reckoned by assuming that the temperature of no variation at 50 feet below the surface is 40° F.

It is to be noted that even these selected records vary a hundred per cent. Very notable variations are found in the same mine or well, and often much difference is found in adjacent records, especially those of artesian wells. Some of these are explainable, but the full meaning of other variations is yet to be found.

Explanations of varying increment.—Certain apparent variations are merely due to *inequalities of topography.* The isogeotherms, or planes of equal underground temperature, do not normally rise and fall with every local irregularity of the surface, but more nearly strike an average. A well on a bluff 500 feet high would probably reach nearly the same temperature at 1000 feet, as a well 500 feet deep in the adjacent valley, giving a gradient twice as great in the one case as in the other.

In interpreting the temperatures of artesian flows, regard must be had to the *depths of rock under which the waters have passed,* as well as the depths at the location of the wells. Darton has found[1] unusually high and varying temperatures in the artesian wells of the Dakotas, some part of which may be due to this cause, though a full explanation of their singular variations is not yet reached.

The permeation and circulation of water affect the temperature in two important ways: (1) wet rocks are better conductors than dry ones, and (2) the convective movement of water is a means of conveying heat from lower to higher horizons. As the circulation of underground water is very unequal, much irregularity of thermal distribution in the upper zones probably arises from this source. The *general* effect of water circulation is to reduce the thermal gradient where the circulation is relatively rapid, as it is near the surface and in the main thoroughfares of circulation, and hence to cause a relatively rapid rise in the gradient just below the zone of effective water influence. Some records conform to this theoretical deduction, but in general it is masked by other influences.

Chemical action, especially oxidation, carbonation, hydration, solution, and precipitation, modify the normal temperature gradient, but how effectively is not well determined. With little doubt the first three mentioned above raise the temperature, while solution and precipitation in some large measure offset each other.[2] The sum-total is probably an appreciable rise in temperature. It has even been conjectured that the heat of volcanic action is due to chemical combination in the lower reaches of water circulation, but this is obviously an over-estimate.

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[1] Am. Jour. of Sci., Vol. V, 1898, p. 161.

[2] Van Hise. Personal communication.

Differences in the conductivity of rock are an obvious source of varying underground temperature gradients. If an outer formation conducts heat more freely than those below, it tends to lower the gradient within itself and to cause a relative rise in the gradient just below. If a lower formation is more conductive than that above, it tends to lower the gradient within itself, and to raise it in the one above, because it carries heat to the outer one faster than the latter carries it away.

The **compression** to which rocks have been subjected affects their temperature. At the surface the variation from this source is chiefly dependent on the lateral thrust suffered.

When allowances are made for all these and other known causes of local variation of temperature, it is still not clear that a uniform average gradient remains as the true conception. If the earth were once a molten spheroid, there would be a strong presumption that, aside from local variations, there would be a normal curve applicable to all regions. On the other hand, if the internal heat has arisen chiefly from compression, and if the compression has varied in different regions, as the inequalities of the surface render probable, there would be no such definite normal curve in the accessible zone of the earth, but rather a varying rate in different regions. In either case, the later movements, compressions and strains of the crust, must modify the original thermal gradients.

Gradients projected.—It is not probable that these gradients, even when corrected for local variations, continue unmodified to the center of the earth. If they did, 1° F. in 60 feet continued to the earth's center would give 348,000° F., and 1° F. in 100 feet would give 209,000° F. It is much more probable that the rates of rise fall away below the superficial zone. If water circulation in the fracture zone is the most efficient agency cooperating with conductivity in the outward conveyance of heat, as seems probable, the gradient in that zone should rise at an abnormal rate, and hence the average gradient in the deeper portions not affected by this circulation should be lower. It will be recalled that the central temperature deduced from an extension of Barus' fusion curve is 136,800° F. (76,000° C.), which, high as it is, gives a lower average gradient than the surface observations. The computations from compression by Lunn, giving a central temperature of 36,000° F. (20,000° C.), imply a still lower average rate, while the convection hypothesis postulates no sensible increase at all below 200 or 300 miles.

The amount of loss of heat.—The amount of loss of interior heat which the earth suffers may be estimated by that which is observed to be passing outward through the rock, or by computing the amount which should be conveyed outwards with the estimated gradients and with the conductivity of rock as determined by experiment. The latter method is usually employed in general problems. Taking the mean thermometric conductivity of rock as 0.0045, the gradient as 1° C. in 30 meters, the average specific heat of rock as 0.5 small calories per cubic centimeter, it is computed that in 100,000,000 years the loss of heat would amount to 45° C. (81° F.) for the whole body of the earth.[2] Tait makes the more conservative estimate of 10° C. (18° F.) in the same period.[3] This is an exceedingly small result, and emphasizes the low conductivity of rock.

Average material of crust (Clarke's tables).[1]		Norm minerals calculated from Clarke's average.		Mineral equivalent (C. I. P. W. system).	Axis.	Linear expansion.	Volume expansion.
SiO_2	58.59	Quartz	11.4	Quartz		+ .00001206	.00003618
Al_2O_3	15.04	Orthoclase	17.2		*a*	+ .00001906	
Fe_2O_3	3.94	Albite	27.3	Anorthite	*b*	− .000002035	
FeO	3.48	Anorthite	17.8		*c*	− .000001495	.00001553
CaO	5.29	Diopside	6.8		*a*	+ .000008125	
MgO	4.49	Hypersthene	10.2	Diopside	*b*	+ .000016963	.0000234
K_2O	2.90	Magnetite and Ilmenite	6.8		*c*	− .000001707	
Na_2O	3.20			Augite (used for hypersthene)	*a*	+ .000013856	
TiO_2	.55	Minor constituents omitted.	2.5				
Minor constituents omitted.	2.52				*b*	+ .00000272	.0000245
			100.00	Magnetite	*c*	+ .00000791	
	100.00					+ .000009540	.00002862

The amount of shrinkage from loss of heat.—To compute the amount of shrinkage for a given amount of cooling, the average coefficient of expansion of rock is required. This has been experimentally determined by several investigators. By combining the determinations of others with his own, T. Mellard Reade found the *linear* coefficient to be

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[1] Bull. 168 U. S. Geol. Surv., p. 14.

[2] Daniell's Physics, p. 407.

[3] Heat. Tait, p. 225.

.000005257 per 1° F., equivalent to .00002838 per 1° C. per *volume.* In this the proportions of the different rocks in the crust were roughly estimated. To secure an independent result from the best available estimate of what constitutes the average rock, W. H. Emmons has reduced Clarke's average of the chemical constituents of the crust to the norm minerals under the new system of Cross, Iddings, Pirsson, and Washington (see p. 454) and made a weighted average of the conductivities of these, as shown in the following table:

	Percentages of norm minerals.	Sp. Gr. of norm minerals.	Volume proportions of norm minerals.	Volume proportions of temp. 1° C. higher.
Quartz	11.4	2.66	4.28	4.2801548504
Feldspars[1]	62.3	2.7	23.07	23.0703582771
Diopside	6.8	3.3	2.06	2.0600482040
Hypersthene	10.2	3.45	2.95	2.9500722750
Magnetite	6.8	5.17	1.3	1.3000372060
Total	97.5		33.66	33.6606708125

Subtracting the stated volume from the volume at a temperature of 1° C. higher, the difference is found to be .0006708125, which divided by the volume gives .0000199, which is the coefficient of expansion of the theoretical, average, surface rock of the earth.

With this coefficient, the radial shrinkage resulting from an average loss of 10° C. (18° F.), (Tait's estimate), is a little over a quarter of a mile (.2572); and for a loss of 45° C. (81° F.), (estimate of Daniell's Physics), a little over a mile (1.1574). The shortening of the circumference for 10° C. loss is 1.6 miles, and for 45° C., 7.27 miles. Computations based on the coefficient of expansion adopted by Reade give 2.35 miles circumferential shortening for a loss of 10° C. and 10.5 miles for a loss of 45° C. In both these cases, the whole contraction is assumed to take a vertical direction, and hence these are maximum results. They are exceedingly small.

Unless there is a very serious error in the estimated rate of thermal loss, or in the coefficients of expansion, cooling would seem to be a very inadequate cause for the shrinkage which the mountain foldings, overthrust faults, and other deformations imply. This inadequacy has been strongly urged by Fisher[1] and by Dutton.[2] In view of the apparent incompetency of external loss of heat, the possibilities of distortion from other causes invite consideration.

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[1] All the feldspars are calculated as anorthite. Augite is used for hypersthene, ilmenite is included with magnetite, and all minerals are calculated as if of the isometric system.

OTHER SOURCES OF DEFORMATION.

Transfer of internal heat.—It is theoretically possible that deformation of the sub-crust may result from the internal transfer of heat without regard to external loss. It has already been shown (p. 539) that under certain possible conditions more heat would flow from the inner parts to higher horizons than would be conveyed through these latter to the surface and there lost, and that, as a result, the temperatures of the inner parts might be falling, while those of the outer parts (except the surface) might be rising. With the more conservative coefficient of expansion previously given, a lowering of the average temperature of the inner half of the earth 500° C. and the raising, by transfer, of the outer half to an equal amount would give a lateral thrust of about 83 miles, which is about the order of magnitude thought to be needed. It is not affirmed that this takes place, but some transfer of this kind is among the theoretical possibilities under the accretion hypothesis. The process could not continue indefinitely; but, for aught that can now be affirmed, it may still be in progress.

Denser aggregation of matter.—As already noted, matter under intense pressure tends to aggregate itself in the forms that give the greatest density. If the earth were built up of heterogeneous matter arranged at haphazard, the material would probably readjust itself more or less, as time went on, into combinations of greater and greater density. This process may be one of the important sources of shrinkage, for an average change of density of 1 percent., affecting the matter of the whole globe, would probably meet all the demands of deformation since the beginning of the Paleozoic period.

Extravasation of lavas.—It is obvious that if lavas are forced out from beneath the crust and spread upon it, a compensating sinking of the crust will follow. This, however, is rather a mode than an ulterior cause, for a cause must be found for the extrusion of the lavas, and this cause may be one of the other agencies recognized, such as a transfer of heat, a reorganization of matter, or a change of pressure. The more practical question, however, relates to its competency. Can the amount of lava that has been extruded have had any very appreciable effect on the descent of the crust? The great Deccan flow is credited with an area of 200,000 square miles, and a thickness of 4000 to 6000 feet. Vast as this is for a lava-flow, it would form a layer only about 5 feet thick when spread over the whole surface of the globe, and hence the sinking to replace it would cause a lateral thrust, on any great circle, of about 31 feet only. It requires a very generous estimate of the lavas poured out between any two great mountain-making periods since the beginning of the well-known stratigraphic series to cause a horizontal thrust of any appreciable part of that involved in mountain-making. The case is different, however, if we go back to the Archean era, in which the proportion of extrusive and intrusive rocks is very high. Very notable distortion may then be assigned to the extravasation of lavas. The outward movement of lava must also be credited with some transfer of heat from lower to higher horizons, and this is probably one of the agencies that have produced the relatively high underground temperatures in the outer part of the earth.

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[1] Physics of the Earth's Crust, Chap. VIII.

[2] Penn Monthly, Philadelphia, May, 1876.

If lavas are thrust into crevices of the crust they contribute to its extension, but causes for the crevices and for the intrusion must be found, and these are probably only expressions of one or another of the more general agencies.

Change in the rate of rotation.—As previously noted, the tide acts as a brake on the rotation of the earth. The oblateness of the present earth is accommodated to its present rate of rotation. It is assumed that such accommodation has always obtained, and that if the rotation has changed, the form of the earth has changed also. Now, the more oblate the spheroid, the larger its surface shell and the less the total force of gravity. Hence if the earth's rotation has diminished, its crust must have shrunk, because the form of the spheroid has become more compact, and the increase of gravity has increased its density. There is at present a water-tide chiefly generated in the southern ocean, and irregularly distributed to more northerly waters. This irregularity interferes with its systematic action as a brake, and its average effects are difficult of estimation. The water-tides of past ages are still more uncertain, as they must have depended on the configuration and continuity of the oceans. There are geological grounds for the belief that the southern ocean was interrupted by land during portions of the past at least, and it is unknown whether there were elsewhere ocean-belts well suited to the generation of large tides. The ocean-tide, therefore, furnishes a very uncertain basis for estimating the retardation of rotation.[1] The theroetical case rests largely on the assumption of an effective body-tide. The earth doubtless has some body-tide, but whether it is sufficiently great to be effective, and whether its position, which depends on its promptness in yielding and in resilience, is favorable to the retardation of rotation, are yet open questions. The existence of an appreciable body-tide has not yet been proved by observation.

G. H. Darwin, assuming that the earth is viscous enough to give a body-tide of appreciable value and of effective position, has deduced a series of former rates of rotation of the earth and has computed the corresponding distances of the moon.[2] C. S. Slichter has shown that the lessening of the area of the surface and the increase of the force of gravity corresponding to these assigned changes of rotation are large, and that if the changes were actually experienced they must have involved much distortion of the crust.[3] These distortions would, however, be of a peculiar nature, and should thereby be detectible, if they were realized; for in passing from a more oblate to a less oblate spheroid, the equatorial belt shrinks, and the polar tracts rise and become more convex. Wrinkles should, therefore, mark the equatorial belts, and tension the high latitudes. Slichter has computed that in a change from a rotation period of 3.82 hours to the present one, the equatorial belt must shorten 1131 miles and the meridional circles lengthen 495 miles. If we take Heim's estimate of the crust-shortening involved in forming the Alps—74 miles—as a standard, the 1131 miles of equatorial shortening would be sufficient for the formation of 15 mountain ranges of Alpine magnitude. If, as some geologists urge, the estimate of mountain folding is too great, the quotient would be still larger. These ranges should run across the equator and be limited to about

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[1] The following conclusion by an eminent authority has come to our notice since this was written:

L'influence des marées océanienes sur la durée du jour est donc tout à fait minime et n'est nullement comparable à l'effet des marées dues à la viscosité et à l'elasticité de la partie solide du globe, effet sur lequel M. Darwin à insisté dans une series de Mémoires du plus haut intérêt. Par H. Poincaré, Bulletin Astronomique, tome XX (June, 1903), p. 223.

[2] On the Secular Changes in the Elements of the Orbit of a Satellite revolving about a Tidally-distorted Planet. Phil. Trans., Roy. Soc., Pt. II, 1880.

[3] Jour. Geol., Vol. VI, 1898, p. 65.

33° N. and S. latitude. The high-latitude tension would be sufficient to cause the earth to gape more than two hundred miles at the poles, if there were simple ideal shrinkage. The amounts and the distribution of thrust and shrinkage are shown in Fig. 453. If the change of rotation were no more than from 14 hours to the present rate, there would still be 52 miles of thrust in the equatorial belt, and 40 miles of shortage in the meridional circles. There are no clear signs of such a remarkable distribution of thrust and tension as this hypothesis requires. Mountains are about as abundant and as strong north of 33°, the neutral line, as south of it, and they extend to high latitudes. The Archean rocks, in which this agency should have been most effective because of their early formation, are crumpled and crushed in the high latitudes much the same as in low latitudes. Furthermore, if there had been appreciable change in the form of the earth to accommodate itself to a slower rotation, the water on the surface, being the most

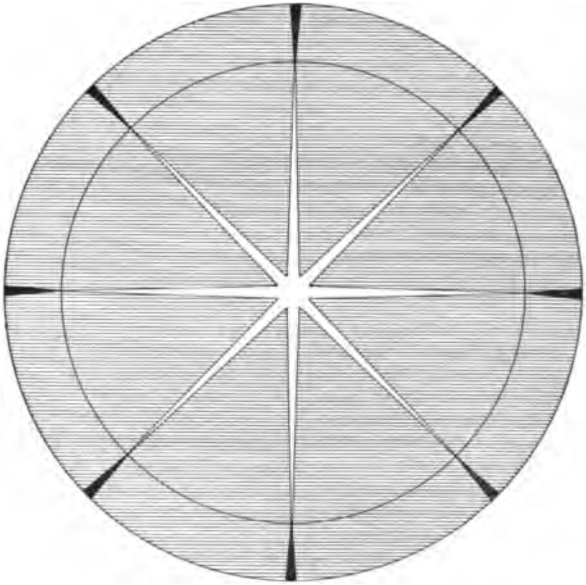

FIG. 453.—Polar projection of the earth's hemisphere showing the theoretical high-latitude tension and low-latitude compression involved in a change of rotation from 3.82 hours to the present rate. The figure is drawn to true scale as seen from a point above the pole, and in consequence the equatorial tract is foreshortened. The black triangles show compression reduced in length by foreshortening; the white show tension in essentially true proportions to the high-latitude areas. The neutral line between the areas of compression and of stretching lies at 33° 20′ latitude.

mobile element, should have gathered toward the poles, and the less mobile solid earth should have protruded about the equator, but the distribution of land and water, present and past, gives no clear evidence of this. The equatorial belt contains a less percentage of land than the area north of it and more than that south of it. It varies but slightly from the average for the whole globe.

While the doctrine of tidal retardation is theoretically sound, and while the relations of the moon to the earth have probably been appreciably affected by tidal action, geological evidence indicates that it has not been sufficiently effective in producing crustal deformations to be clearly detected by its own distinctive results. This may be due (1) to the fact that there are compensating agencies that tend to acceleration of rotation, and (2) to the probable fact that the central rigidity of the earth is too high to give a very effective body-tide. Hence the process of retardation may have been too slow to have been geologically appreciable in the known period. The recent estimates of the effective rigidity of the earth are greater than former ones, and they may need to be modified yet further in the same direction.

Distribution of rigidity.—An important consideration in this connection is the distribution of interior rigidity. It is certain that the rigidity of the outermost part, taken as a mass, is somewhat less than that of rock of an average surface type, for it is fissured, and there is no reason to suppose that the rigidity of the rock next below the fissure zone rises at once to the rigidity of steel, and hence if the average rigidity of the whole earth is equal to that of steel, a portion of the interior must have a rigidity much higher than steel. There is probably some law of increase from surface to center, and there are theoretical grounds for thinking that it is in some way connected with the laws of pressure, density, compressibility, and temperature. All of these factors probably affect rigidity, but in different ways. The modulus of rigidity of steel is about 770×10^6 grms. per sq. cm. Milne and Gray[1] found that of granite to be 128×10^6. The ratio of the rigidity of steel to that of rock is, therefore, about 6 : 1. If it be assumed that the rigidity increases in depth directly as the density, the rigidity will nowhere reach that of steel, being only about two-thirds as much at the center.

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[1] Quar. Jour. Geol. Soc., Vol. 39, 1883, p. 140. Everett (Units and Physical Constants) gives 837×10^6 for steel, but as the modulus for granite seems low, we have taken the lower estimate for steel to avoid exaggerating the ratio between them.

If it be assumed that the rigidity increases as the squares of the density ratios, the following values are obtained:

Distances from center in terms of radius.	Densities under Laplace's law.	Density ratios.	Density ratios squared.	Deduced rigidities.
1.00	2.8	1	1	0.16 Steel
.75	5.7	2	4	0.6 "
.50	8.39	3	9	1.5 "
.25	10.27	3.7	13.7	2.3 "
.00	10.95	3.9	15.2	2.5 "

These values seem fairly consistent with the apparent requirements of the case.

If the distribution of rigidity were of this nature, the *average* rigidity would be much less than that of steel, for more than half the volume lies in the outer division, between 1.00 and .75 radius, and yet the effective resistance to tidal deformation would be high, for, according to G. H. Darwin,[1] the tidal stress-differences are eight times as great in the center as at the surface. The rigidity would, therefore, be distributed so as to be much more effective in resistance than if it were uniform. The suggestion arises here that the tidal stresses and other analogous stresses arising from astronomical sources may be in themselves the causes of some such distribution of rigidity as this. The tidal stresses are rhythmical and give rise to a kind of kneading of the body of the earth, small in measure to be sure, but persistent and rapidly recurrent. Since these stress-differences at the center are eight times those at the surface, and since also the gravitative stress at the center is 3,000,000 times that at the surface, there is a series of persistently recurring stress-differences, greatest at the center and declining outwards, superposed on enormous static stresses, also intensest at the center and declining outwards. Now, if the earth material were once made up of a mixture of minerals of different fusibility, some of which became more mobile (whether fluid or viscous) than others under the rising temperature of the interior, it seems that the more mobile portion must have tended to move from the regions of greater stress-differences to those of lesser stress-differences. The persistence and the rhythmical nature of the tidal stress-differences seem well suited to aid the mobile parts in gradually working their way outwards. At the same time the more solid and resistant portions should remain behind, and thus come to constitute the dominant material of the central regions where stress-differences were greatest, and so, as it were, concentrate rigidity there. The process may still be in action.

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[1] Nat. Phil. Thompson & Tait, Vol. II, p. 424, 1890.

If it be assumed that the rhythmical stresses have thus developed a resistance to deformation proportional to their intensity, we may combine this with density to form the basis of another hypothetical distribution of rigidity, as follows:

Distances from center in terms of radius.	Densities under Laplace's law.	Density ratios.	Ratios adjusted to stress-differences. (1 : 8)	Deduced rigidities.
1.00	2.8	1	1	0.16 Steel
.75	5.7	2	3.5	0.58 "
.50	8.39	3	5.4	0.90 "
.25	10.27	3.7	7	1.16 "
.00	10.95	3.9	8	1.33 "

The average rigidity is here also much less than that of steel, but its distribution is such as to render it ideally fitted to resist tidal distortion.

These hypothetical distributions of rigidity have no claims to special value in themselves, for the grounds on which they are based are quite inadequate, but they are not without importance in giving tangible form to considerations that bear vitally not only on tidal problems, but on many others connected with the internal constitution and dynamics of the earth.

Sphericity as a factor in deformation.

It is obvious that if the earth shrinks, its crust must become too large for the reduced spheroid, and must be compressed or distorted to fit the new form. The amount of distortion required for any given shrinkage is easily computed from the ratio of the radius to the circumference of a sphere, which is approximately 1:6.28. If, for example, the radius shortens 5 miles, each great circle must on the average be compressed, wrinkled, or otherwise distorted to the extent of about 31 miles, or, in reversed application, if the mountain foldings on any great circle together show a shortening of 100 miles, the appropriate radial shortening is 16 miles. The ratio of 1:6+ furnishes a convenient check on hypotheses that assign specific thrusts to specific sinkings of adjacent segments. A segment 3000 miles across, for example, such as the bottom of the North Atlantic basin, sinking three miles, about the full depth of the basin, would give a lateral thrust of about 2.2

miles, a little over a mile on each side, a trivial amount compared with the foldings on the adjacent continental borders.

The influence of the domed form of the surface.—Because of the spheroidal form of the earth, each portion of the crust is ideally an arch or dome. When broad areas like the continents are considered, it is the dome rather than the arch that is involved, and in this the thrust is ideally toward all parts of the periphery. It is probably for this reason that mountain ranges so often follow curved or angulated lines, or outline rude triangles or polygons. The sigmoidal courses of the ranges of southern Europe, the looped chains of the eastern border of Asia, and the curved ranges of the Antillean region, are notable examples. The border ranges of the Americas, of the Thibetan plateau, and of other great segments, illustrate the polygonal tendency. The general distribution of the great ranges is such that a nearly equal portion of crustal crumpling is thrown across each great circle, as theory demands. The common generalization that mountain ranges run chiefly in oblique directions, as northeast-southwest, northwest-southeast, is but a partial view of the more general fact that the lines of distortion must lie in all directions to accommodate the old crust to the new geoid, if there be equable contraction in all parts.

Theoretical strength of domes of earth-dimensions.—As the domed form of the crust has played an important part in theories of deformation, it is important to form quantitative conceptions of the strength of ideal domes having the figure and dimensions of segments of the earth's crust. According to Hoskins,[1] a dome corresponding perfectly to the sphericity of the earth, formed of firm crystalline rock of the high crushing strength of 25,000 pounds to the square inch, and having a weight of 180 pounds to the cubic foot, would, if unsupported below, sustain *only* $\frac{1}{528}$ *of its own weight.*[2] This result is essentially independent of the extent of the dome, and also of its thickness, provided the former is continental and the latter does not exceed a small fraction of the earth's radius. If this ideal case be modified by supposing the central part of the spherical dome to rise above the average surface,

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[1] Computations made at the request of the authors. See also Fisher, Physics of the Earth's Crust, p. 36.

[2] Of like import is the statement of Woodward—"If the crust of the earth were self-supporting, its crushing strength would have to be about thirty times that of the best cast steel, or five hundred to one thousand times that of granite." Mathematical Theories of the Earth, Proc. Am. Assoc. for Adv. Sci., 1889, p. 49.

the supporting power will not be materially changed unless the central elevation is a considerable fraction of the radius of the dome. Assuming a central elevation of two miles—to represent the protrusion of the continental segments—the results for domes of different horizontal extent are as follows:[1]

THEORETICAL STRENGTH OF IDEAL DOMES ARCHED TWO MILES ABOVE THE AVERAGE SURFACE OF THE SPHERE.

Diameter of given dome arched 2 miles above sphere.	Multiplier of 1/525 i.e. the supporting proportion of a spherical dome.	Proportion of its own weight sustained by given dome arched 2 miles above sphere.
3,000 miles	1.006	1/522
400 "	1.396	1/376
240 "	2.11	1/249
160 "	3.49	1/150
80 "	10.97	1/48

From this table it will be seen that for domes of continental dimensions the supporting strength *equals only a very small fraction of the dome's own weight.* Increasing the thickness of the shell increases its actual supporting power, but the proportion is somewhat less when the whole sphere is concerned. The problem has not been worked out for domes of limited extent. For rough estimates, where the dimensions of the dome are of continental magnitude, each mile of thickness may be taken as supporting a layer of about 10 feet of its own material. If the hypothetical level of no stress be placed at 8 miles depth, the shell above this, by reason of its domed shape, could relieve its own pressure on that below to an amount equal only to the weight of about 80 feet of rock over its surface, even if its form and structure were ideal. If the shell were thick enough (817 miles) to embrace one-half the volume of the earth, its supporting power would be a little more than the weight of one and one-half miles of rock. As the radius of the earth is less than 4000 miles, the extreme supporting power reckoned on this basis would be only about 8 miles of rock-depth. It is interesting, if not significant, to observe that this depth barely reaches the minimum shrinkage that will serve, according to current estimates, to account for the crustal shortening of the great mountain-making periods. It is as if the shrinkage stresses accumulated to the full extent of the stress-resisting power of the whole sphere, and then collapsed. It is not safe, however, to give much weight to this coincidence, for higher densities and probably higher resistances to distortion come into play in the deeper horizons. If these resistances are proportional to the higher densities of the interior, the deductions would remain the same. If the effective rigidity of the earth as a whole is that of steel, as deduced by Kelvin and Darwin from tidal and other observations, or twice that of steel, as inferred by Milne from the transmission of seismic vibrations, the supporting power of the body of the earth dependent on its sphericity would be appreciably higher.

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[1] It is assumed that the direction of the supporting thrust at the periphery of the dome is at every point parallel to the tangent to the domed surface. This is justified by symmetry in the case of a shell conforming to the sphericity of the earth, and in the other cases it would seem to be as favorable an assumption in the direction of high supporting capacity as can reasonably be made.

It would seem clear from the foregoing considerations that something more than the mere crust of the earth has been involved in the great deformations. Indeed it is not clear that the fullest resources of stress-accumulation which the spheroidal form of the earth affords are sufficient to meet the demands of the problem, unless the rigidity of the earth be taken at a much higher value than that of surface-rock, and this is perhaps an additional argument for the high rigidities inferred from tides and seismic waves.

In view of the doubtful competency of even the thickest segments to accumulate the requisite stresses, there is need to consider modes of differential stress-accumulation other than those dependent on sphericity.

Stress-accumulation independent of sphericity.—The principle of the dome is brought into play whenever an interior shell shrinks away, or tends to shrink away, from an outer one which does not shrink. In this case, there is a free outer surface and a more or less unsupported under surface toward which motion is possible. The dome may, therefore, yield by crushing or by contortion. The computations given above are for cases of this kind. But where the thickness becomes great and the dome involves a large part or even all of a sector of the earth, freedom of motion beneath is small, and to readjust the matter to a new form, strains must be developed widely throughout the sector, and must involve regions where the pressure is extremely great on all sides, and crushing in the usual sense impossible. Assuming the correctness of the modern doctrine that such pressure increases rigidity, instead of the older doctrine that it gives plasticity, it becomes reasonable to assume that stress-differences would be distributed throughout the mass, and bring into play a large portion of its stress-accumulating competency. When the mass yielded, it would not be by crushing, but by "flowage," which would be more or less general throughout the mass. It might, however, be partially concentrated, as, for example, on the borders of sectors of different specific gravity.

Stress-differences may arise from physical changes within the rock itself. Whenever there is a re-aggregation of matter, or a change of any kind which involves change of volume, a change of stress is liable to be involved. It may be of the nature of relief or of intensification. In an earth built up by the haphazard infall of matter, a very heterogeneous mass must result, and the subsequent changes may be supposed to be intimately distributed through the mass, being slight at any point, but present at innumerable points. An immeasurable number of small stress-differences may, therefore, be developed throughout the mass. Until these overmatch the effective strength of the mass, they may continue to accumulate. These are not necessarily connected with stresses that arise from sphericity, and may work more or less independently of them. It is not improbable that the great stress-accumulating power of the globe finds an essential part of its explanation in supplemental considerations of this kind, and not wholly in its spheroidal form.

The actual configuration of the surface.—The foregoing computations relative to the power of shells of the earth to sustain pressures are based on ideal forms and structures that are not realized in fact. How far the earth fails to conform to these conditions must now be considered. When compared with the earth as a whole, the inequalities of its surface are trivial. If the great dynamic forces acted through the whole or the larger part of the body of the earth, the configuration of the surface can be supposed to have done little more than influence the location of the surface deformations and their special phases. But if the forces were limited to a crust of moderate thickness, the configuration of the surface is a matter of radical importance.

Concave tracts.—There is need, therefore, to inquire if any considerable breadth of the crust is outwardly plane or concave, for the principle of the dome is obviously not applicable to a plane or concave surface. To be a source of fatal weakness, the concavity must be broad enough to cause the planes of equal cooling, the isogeotherms, to be concave to considerable depths. For example, if the hypothetical level of no stress is eight miles below the surface, as computed on certain assumptions, the concave portion must be so broad that the isogeotherms will also be concave outward at something near that depth; in other words, the main part of the zone of thrust must be concave. A narrow concavity at the surface, such as an ordinary valley in a portion of the crust that has the average convexity, would not seriously depress the isogeotherms, or affect the zone of thrust, but a valley several times eight miles (level of no stress) in breadth would. For inspecting the surface of the earth in this regard, it is convenient to know what amounts of fall below the level surface give a true plane for given distances. These are shown in the following table:[1]

Length of arc in miles.	Length of normal to chord at middle point in		Average fall of true plane from level plane per mile, in feet. Greater fall gives concavity.
	Feet.	Fathoms.	
25	100.3	16.7	8.
50	432.	72.	17.3
75	913.4	152.2	24.3
100	1,684.	280.7	33.7
150	3,748.8	624.8	49.9
200	6,674.	1,112.3	66.7
250	10,369.9	1,728.3	82.9
300	14,942.	2,490.3	99.6
400	26,664.	4,444.	133.3
500	41,659.	6,943.	166.6

Applying these criteria to the surface of the lithosphere, *it is found that concave tracts from* 100 *to* 300 *miles in breadth are not uncommon.* The more notable of these are shown in black on the accompanying map, Fig. 454, and two typical ones are shown in cross-section in Figs. 455 and 456. It is to be observed that concave tracts border the continents very generally. They are connected with the descent from the continental shelf to the abysmal basins, and are unsymmetrical. Notable concavities are found in some of the great valleys on the continental platforms. The basins of Lake Superior, Michigan, Huron, and Ontario are in part concave; so are Puget Sound, the Adriatic, and the Dead Sea; so also are the valleys of California, of the Po, and of the Ganges, when the adjacent mountains are included. Some of

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[1] Prepared at the authors' request by W. H. Emmons.

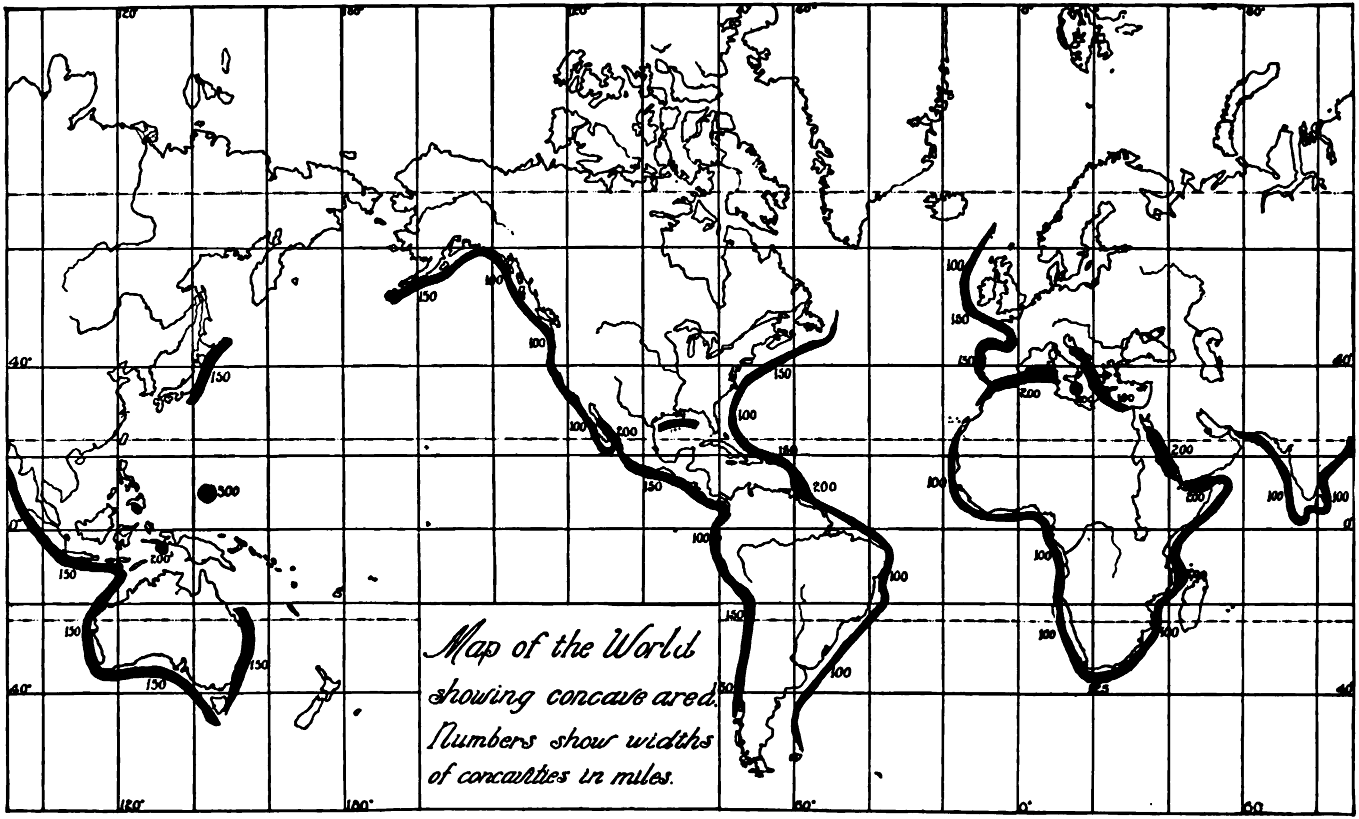

Fig. 454.—Map of the world, showing in black the chief submarine concavities of the lithosphere. (Prepared by W. H. Emmons.)
(Page 586.)

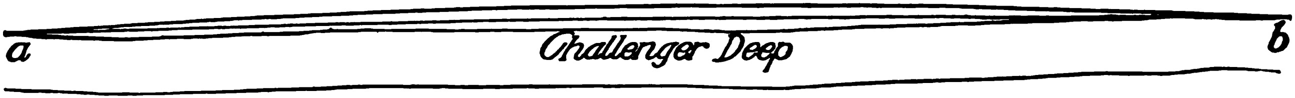

FIG. 455.—Section of the Challenger Deep from an island on the Caroline plateau, *a*, to an island on the Ladrone plateau, *b*, *drawn to a true scale*, showing the real concavity of the surface of the lithosphere for a breadth of 300 miles. The upper line represents the sea surface, a natural level. The next line below represents a true plane, eliminating the curvature of the sea surface. The third line represents the bottom of the deep. By comparison with the line above, its true concavity may be seen. The lowest line represents an isogeotherm at about 10 miles below the surface; i.e. appreciably below "the level of no stress," as usually computed, showing that the whole thrust zone is concave outwards, if it is limited to surface cooling as usually computed. (Prepared by W. II. Emmons.)

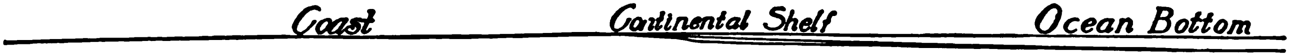

FIG. 456.—Section through the Atlantic coastal plain, the continental shelf, and a portion of the abysmal bottom, *drawn to a true scale*, showing that the surface of the lithosphere drops below a true plane tangent to the continental shelf and the ocean-bottom. The upper line represents the surface of the coastal plain at the left and of the ocean at the right. The lower line represents the sea-bottom, and the middle line a true plane tangent to the shelf and the sea-bottom. The breadth of the concave tract varies from 100 to 150 miles. (Prepared by W. H. Emmons.)

the "deeps" of the bottom of the ocean are notably concave. Fig. 455, a cross-section of the Challenger Deep, *drawn to true scale and convexity,* shows the nature of the phenomenon. The breadth is here 300 miles, and the depression below a true plane is 11,400 feet. The lower line of the figure shows the approximate position and form of the normal isogeotherm about ten miles below the surface. Assuming equal conductivity in all parts, it is clear that the isogeotherms must be concave upwards for a considerable distance below ten miles. Unless the shell of thrust is much more than ten miles thick, these concave portions should yield as fast as cooling below them permits, and *no stresses arising from convexity could be accumulated.*

These concavities of surface are so extensive and so widely distributed over the globe that *no part of the outer shell can be supposed to be capable of accumulating notable stresses* unless rigidly attached to the earth-body below. In other words, so far as sphericity is concerned, the crust must ease all its stresses nearly as fast as they accumulate, if, as usually assumed, it rests on a contracting or mobile substratum.

Surface cooling under these conditions should give only feeble thrusts, developed and eased nearly constantly. Such movements should be admirably adapted to give those gentle, nearly constant subsidences that furnish the nice adjustments of water-depth required for the accumulation of thick strata in shallow water, and those slow upward warpings that renew the feeding-grounds of erosion, the necessary complement of the deposition. These gentle, nearly constant movements mark every stage of geological history, and constitute one of its greatest though least obtrusive features. *But if superficial stresses arising in this way are eased in producing these effects, they cannot accumulate to cause the great periodic movements.*

Even where the crust is not concave, it is so warped and so traversed by folds and fault-planes that its resistance to thrust is relatively low, and it should, therefore, warp easily and at many points, if the thrust be confined to a superficial crust.

General conclusion.—When to the weakness of the crust, as computed under ideal conditions, there is added the weakness inherent in these concave and warped tracts, the conclusion seems imperative that while the crust is the pliant subject of minor and nearly constant warpings, such as are everywhere implied in the stratigraphic series, it is wholly incompetent to be the medium of those great deformations which occur at long intervals and mark off the great eras of geologic history. These great deformations apparently involve the whole, or a large part, of the body of the earth, and seem to require a very high state of effective rigidity.

General references on crustal movements.—Babbage, Jour. Geol. Soc., Vol. III (1834), p. 206; Lyell, Principles of Geology, Vol. II, p. 235; Mallet, Phil. Trans. (1873), p. 205; Reade, Origin of Mountain Ranges, and Evolution of Earth Structure; Fisher, Physics of the Earth's Crust; Dutton, Greater Problems of Physical Geology, Bull. Phil. Soc. of Washington, Vol. XI, p. 52, also Amer. Jour. of Sci., Vol. VIII (1874), p. 121, and Geology of the High Plateaus of Utah (1880); Jamieson, Quar. Jour. Geol. Soc. (1882), and Geol. Mag. (1882), pp. 400 and 526; Heim, Mechanismus der Gebirgsbildung; Marjerie and Heim, Les Dislocations de l'Écorce terrestre (1888); Shaler, Proc. Boston Soc. Nat. Hist., Vol. XVII, p. 288; Dana, Manual of Geol., 4th ed., p. 345 et seq.; Woodward, Mathematical Theories of the Earth, Smithsonian Rept. for 1890, p. 196; Willis, The Mechanics of the Appalachian Structures, 13th Ann. Rept. U. S. Geol. Surv., Pt. II (1893), pp. 211–282; LeConte, Theories of Mountain Origin, Jour. Geol. Vol. I (1893), p. 542; Gilbert, Jour. Geol., Vol. III (1895), p. 333, and Bull. Phil. Soc. of Washington, Vol. XIII (1895), p. 31; Van Hise, Earth Movements, Trans. Wis. Acad. Sci., Arts and Let., Vol. II (1898), pp. 512–514; Estimates and Causes of Crustal Shortening, Jour Geol., Vol. VI (1898), pp. 29–31; Relations of Rock Flowage to Mountain Making, Mon. XLVII, U. S. Geol. Surv. (1904), pp. 924–931; A. Geikie, Text-book of Geology, 4th ed., pp. 672–702.

CHAPTER X.

THE EXTRUSIVE PROCESSES.

Outward movements.—In the preceding chapters movements toward the center have been considered. The complementary processes of outward movement now invite attention. Without doubt these are mainly but a resultant of the centripetal actions. For each pound of material moved outwards an equivalent is quite surely moved inwards. Notwithstanding this, the outward movements have a peculiar nature of their own, and serve a function of radical importance in the economy of the globe. Some minor phases have been incidentally considered, such as the upward flow of springs and deep-seated waters, but here the descending and ascending factors are alike, and are closely and obviously connected.

VULCANISM.

The great example of ascensive action is the movement of fluid rock from the interior outwards. The term vulcanism will be used to embrace not only volcanic phenomena in the narrower sense, but all outward forcing of molten material, whether strictly extrusive or merely ascensive.

The philosophy of this ascensive action, taken as a whole, is simple. In the effort at concentration under the powerful action of the earth's gravity, the material of high specific gravity is urged more strongly toward the center, volume for volume, than that of less specific gravity, and as gravity is perpetually active, it follows that whenever any movement, molecular or molar, takes place which permits a readjustment of the positions of the two kinds of matter, the heavier sinks toward the center and the lighter rises, or at least tends to do so. So also where there are stress-differences, the mobile matter tends to flow from the regions of greater stress toward those of lesser stress. In so far as any portion of the interior becomes liquid, it is free to move up or down according to the balance of stress brought to bear upon it, and adapts itself to any line of least resistance available to it. As a natural result, therefore, the portion of the interior which becomes fluid most largely participates in the outward movement. In so far as molecular action permits a readjustment of material, there is a tendency, even in the solid state, for the lighter material to move upwards and the heavier downwards, and for the more stressed portions to move toward points of less stress; but this takes place with extreme slowness. In so far as the materials of the interior diffuse themselves through each other, the same laws hold good, but they are modified by the special principles that control diffusion. The outward diffusion of interior gases may be a factor of appreciable importance, but this cannot be affirmed at present.

Phases of vulcanism.—The forcing of fluid rock outward assumes two general phases, which, however, merge into each other; and these main phases take on various sub-phases. The first phase embraces those outward movements of fluid rock which do not reach the surface. The lavas, after ascending to the vicinity of the surface, intrude themselves into the outer formations of the earth and congeal underground (plutonic). The second phase embraces those outward movements in which the fluid rock reaches the surface and gives rise to eruptive phenomena (volcanic). The first is *intrusive*, the second *extrusive;* the first constitutes *irruptions*, the second *eruptions*.[1] The fundamental nature of the two is the same, but the extrusions usually take on special phases because of the relief of pressure at the surface of the earth, and because of the action of surface-waters in contact with the heated lavas. Just where the lavas come from, and how they find their way through the deep-lying compact zone below the zone of fracture, may better be considered later. When they reach the zone of fracture, they usually either take advantage of fissures already formed, or force passageways for themselves by fracture. There is little evidence that they bore their way through the rocks by melting, though they appear to round out their channels in some way into pipes, ducts, and other tubular forms when they flow through them for long periods of time.

1. *Intrusions.*

Fluid rock forced into fissures and solidified there forms *dikes;* forced into chimney-like passages, it forms *pipes* or *plugs;* insinuated between beds, it forms *sills;* bunched under strata so as to arch them upwards, it forms *laccoliths;* massed in great aggregations underground, it constitutes *batholiths*, as already described (pp. 394 and 500). Lavas sometimes crowd aside the adjacent rocks so far as to cause them to take a concentric form about the intruded mass. This is not uncommon in the oldest formations, and is probably not infrequent in the deeper horizons where the pressures are very great. Some part of this may, however, be due to later deformations. Nearer the surface, usually, the beds are merely lifted as in forming the sills, or are bowed upwards, as in the laccoliths, or faulted as in *bysmaliths* (p. 500).

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[1] The terms are here used in their narrow technical sense. Extrusion is also used in a broad generic sense to indicate the whole process of outward movement.

The heating action on the adjacent rock varies greatly with the mass and temperature of the intruded lava. Thin dikes and sills often produce little effect, while greater and hotter masses notably metamorphose the adjacent rock. In some cases marked effects are due to a thin stream of lava flowing through a fissure for a long period, and so maintaining a high temperature. In the least effective cases, the adjacent rock usually shows some signs of baking. In the marked cases, there is more or less new crystallization. The surrounding rock commonly shows some evidence of material derived from the lavas; less often the lava shows some evidence of having received material from the adjacent rock. But since the lavas do not usually bore their way through the strata in the zone of fracture, nor melt the adjacent rock, the constitution of the lavas is not appreciably changed by the kinds of rock which they penetrate. On the other hand, the intrusions often show the effects of rather rapid cooling by contact with the adjacent rock, (*a*) by a less coarse crystallization near the rock-walls, and sometimes (*b*) in a segregation of the material.

2. *Extrusions.*

When molten rock is forced to the surface it gives rise to the most intense and impressive of all geological phenomena. The energies acquired in the interior under great compression here find sudden relief. Occluded gases often expand with extreme violence, hurling portions of the lavas to great heights and shattering them into fragments constituting "smoke," ash, cinders, bombs, and other pyroclastic material. Much of the explosive violence of volcanoes has been attributed to the contact of surface-waters with the hot rising lava, but the function of this kind of action has probably been exaggerated.

There are two phases of extrusion often quite strongly contrasted. The one is explosive ejection, often attended with great violence; the other, a quiet out-welling of the lava, with little more than ebullition. More or less closely related to these differences are two classes of conduits, (*a*) the one, great fissures, out of which the lava pours in great volume and spreads forth over wide tracts, often in broad thin sheets; (*b*) the other, restricted openings, often pipes, ducts, or limited fissures, from which the extrusion is usually much less abundant, and hence it more largely congeals near the orifice, forming cones. Flows from the former constitute massive eruptions; those from the latter, the more familiar volcanic eruptions. There is no radical difference between them, and the two classes blend. The extent of the spreading of lava into thin sheets is due more to the mass and the fluidity than to the form of the outlet. The stupendous outflows of certain geologic periods appear to have issued mainly from extended fissures, doubtless because these better accommodated the outbursting floods.

a. **Fissure eruptions.**—The chief known fissure eruptions of recent times are the vast basaltic floods of Iceland. Most of the eruptions of historic times are of the volcanic type; but at certain times in the past there were prodigious outpourings, flow following flow until layers thousands of feet thick covering thousands of square miles were built up. One of these occurred in Tertiary times in Idaho, Oregon, and Washington, where some 200,000 square miles were covered with sheets of lava, aggregating in places 2000 feet or more in thickness. Earlier than this, in Cretaceous times, there were enormous flows on the Deccan plateau of India, covering a like area to a depth of 4000 to 6000 feet. Still earlier than this, in Keweenawan times, an even more prolonged succession of lava-flows covered nearly all the area of the Lake Superior basin, and extended beyond it, and built up a series of almost incredible thickness, the estimates reaching 15,000 to 25,000 feet. In these cases there is little evidence of explosive or other violent action. There are few beds of ash, cinders, and similar pyroclastic material. The inference is, therefore, that the lavas welled out rather quietly and spread themselves rather fluently over the surrounding country. For the most part these wide-spreading flows are composed of basic material, which is more easily fusible and more highly fluent at a given temperature than the acidic lavas. The latter are more disposed to form thick embossments near the point of extrusion.

Massive outflows of this class constitute by far the greatest phenomena of the extrusive type, though they are not now the dominant type. It has been sometimes thought that the more local volcanic type of extrusion followed the more massive fissure type as a phase of decline; but this has not been substantiated.

b. **Volcanic eruptions.**—In the types of eruption prevailing at the present time, the lavas are forced out through ducts or perhaps short fissures or sections of fissures, and build up cones about the vents, the eruptive action maintaining craters in the centers of the cones. The essential feature of a volcano is the issuance of hot rock and gas from a local vent. A mountain is the usual result, but the mountain is secondary and not usually present in the first stages; the localized eruption is the primary and necessary factor. The amount of rock

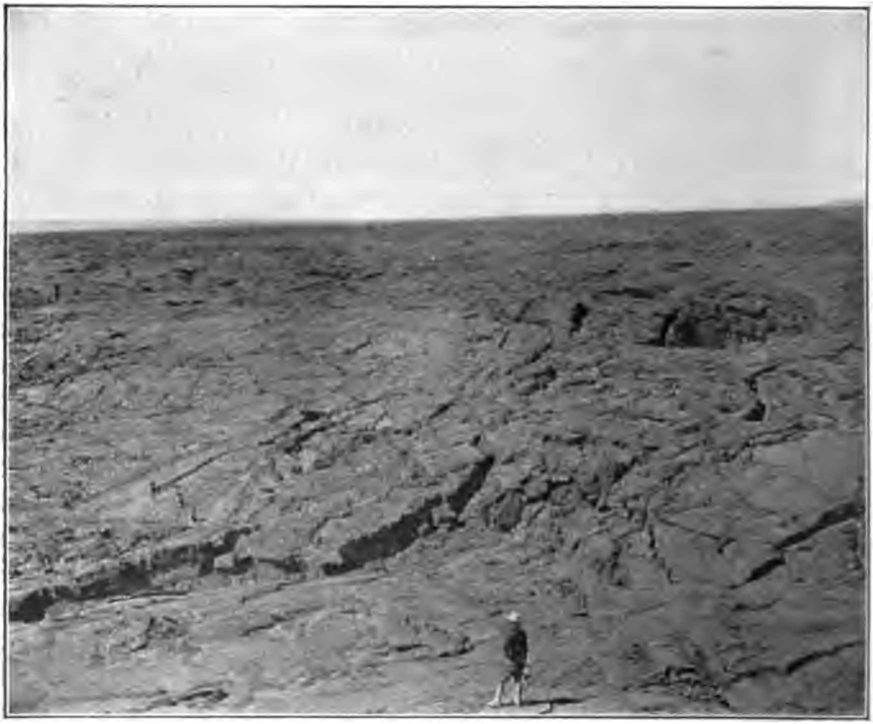

FIG. 457.—Lava-flow near the Jordan craters, Malheur Co., Oregon. Though not of the gigantic order, it illustrates the general aspect of massive lava-flows. (Russell, U. S. Geol. Surv.)

matter ejected is not necessarily great. Compared to the massive extrusions of fissure eruptions, it is usually rather trivial; but the volcano makes up in demonstrativeness what it lacks in massiveness of product.

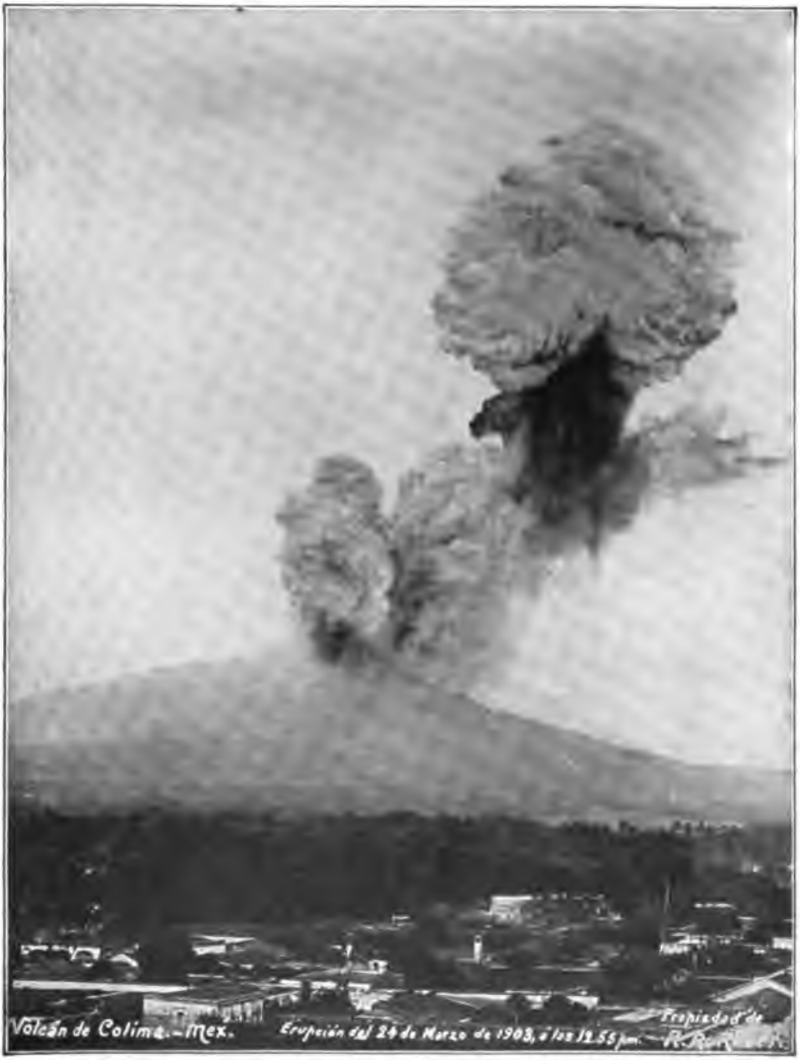

FIG. 458.—The volcano Colima, Mexico, in eruption. March 24, 1903. (José Maria Arreola, per Frederick Starr.)

c. **Intermediate phenomena.**—On the border-line between the intrusive and the extrusive phenomena there are special cases of interest. There appear to be certain instances in which the intrusion comes so near the surface as to develop explosive phenomena without the extrusion of lava. From the nature of the case this is an interpretation rather than a demonstration. It is certain, however, that occasional violent explosions take place where no lava comes in sight. This sometimes occurs in old volcanic formations, and sometimes in regions of undisturbed horizontal strata. In the former case the phenomena may be due to the intrusion of a fresh tongue of lava below, or it may be due to the penetration of surface-waters to hot rocks that have remained uncooled from previous volcanic action, and the development, by such contact, of a volume of confined steam sufficient to produce the explosion. A case of this doubtful kind occurred at Bandai-San in Japan in 1888, where there was a sudden and violent explosion which blew away a considerable part of the side of a volcanic mountain which had not been in eruption for at least a thousand years. The mass and violence of the exploded material was such as to fill the air with ashes and débris in a fashion altogether similar to a typical volcanic eruption. A large tract of adjacent country was devastated, and many lives lost. The whole action, however, was concentrated in the initial explosion, and within a few hours the cloud of ashes had disappeared and the phenomenon was ended. An examination of the disrupted area revealed no signs of liquid lava.

An example of the latter class is Coon Butte in Arizona.[1] This consists of a rim of fragmental material encircling a crater-like pit from which the fragments were obviously forced by violent explosion. The pit is in ordinary sedimentary strata, and the material of the rim is composed of the disrupted fragments of the sedimentary rock ejected from the pit. There are no signs of igneous material, but there was igneous action in the vicinity. Fragments of a meteorite were found on the rim and in the vicinity, but this association appears to be accidental. Computation shows that the volume of the material of the rim closely matches the size of the pit. The source of the explosion is not demonstrable, and it may be an error to connect it with an intrusion of lava below; but since intrusions rise to various degrees of nearness to the surface, and in innumerable cases reach the surface, there is every reason to entertain the concep-

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[1] Gilbert. 14th Ann. Rept. U. S. Geol. Surv., Pt. I, p. 187.

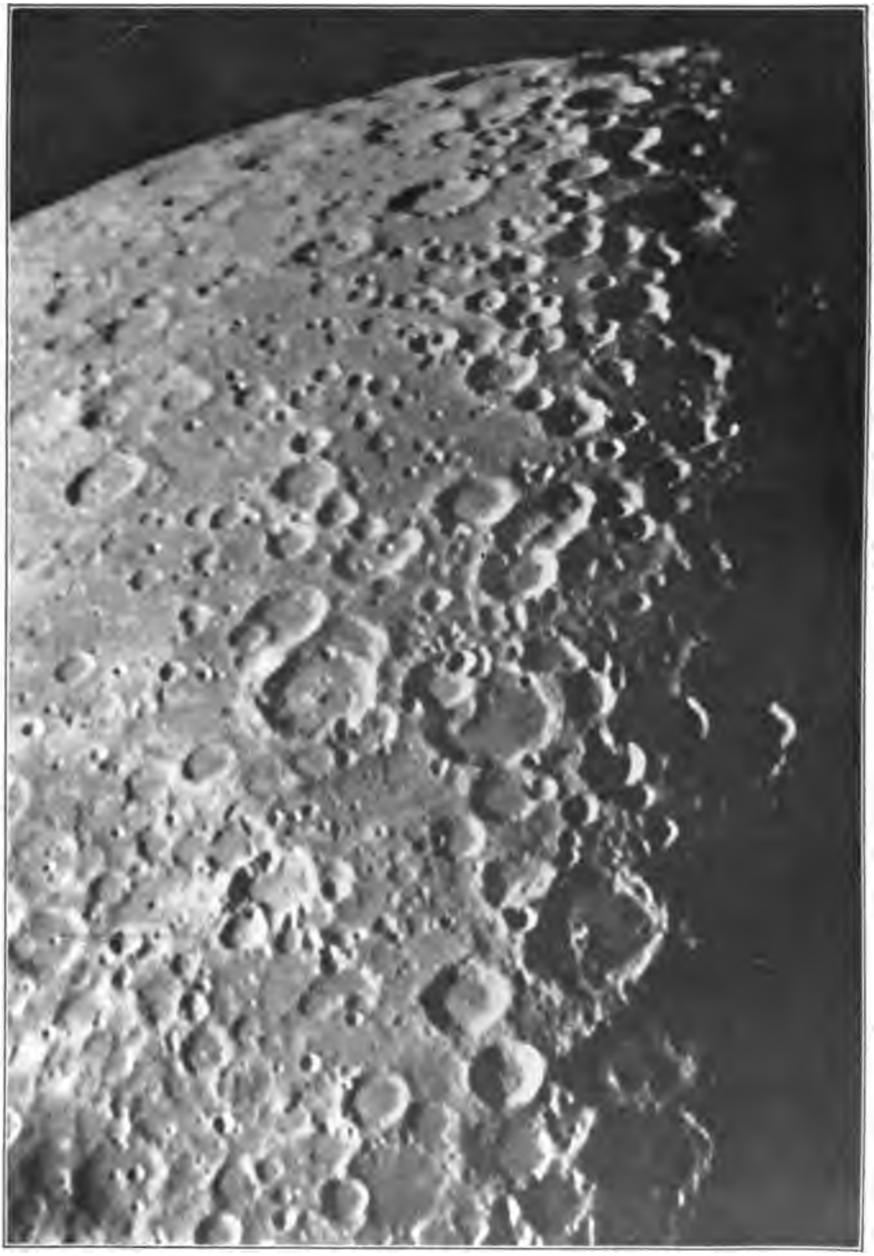

Fig. 459.—Photograph of a portion of the moon taken at Lick Observatory.
(Page 597.)

tion of a class of intrusions which develop explosive phenomena by close approach to the surface, without actually reaching it.

Lunar craters.—There are grounds for thinking that the remarkable craters of the moon, assuming that they are truly volcanic,[1] may belong to this class, for they are very similar to the Coon Butte pit. The capacities of the lunar craters, so far as they can be estimated, seem to equal, if they do not in many cases exceed, the volume of matter in their rims. They do not appear usually to be great cones of accumulated material with relatively small craters, like the typical products of terrestrial volcanoes. Besides, there are no clear evidences of lava-streams. The radiating tracts once interpreted as such have been shown by increased telescopic power and the resources of photography to be at least something other than lava-streams. They are vaguely defined tracts which run over heights and depths indifferently, and are plausibly interpreted as lines of débris projected to extraordinary distances because of the absence of a lunar atmosphere, and because of the low force of the moon's gravity. Since the moon now has no appreciable atmosphere or surface-waters, and since it is doubtful whether it ever possessed either on account of its probable inability to hold atmospheric gases or the vapor of water in the form of an envelope about it, owing to its low gravity, there is reason to suppose that the external matter of the moon derived from the explosions of the multitude of lunar volcanoes would remain in a loose, incoherent condition, from the absence of dissolving and cementing agencies. It is reasonable to suppose that lava-tongues arising from the deeper interior would have a higher specific gravity, even in their heated condition, than this porous covering of the moon, and that therefore they would almost universally become intrusions rather than extrusions, or at most they would not rise beyond the bottom of the craters they had produced by explosion. This seems to furnish at least a plausible explanation of the prevailing differences between the large lunar craters encircled by mere rims and the much smaller terrestrial craters seated in relatively large cones.

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[1] Gilbert, after a careful study of the moon's topography, has suggested that the lunar pits may be indentations produced by infalling meteorites or planetoids, and has shown by experiment that pits of a similar type, with similar central cones, can be produced by impact. The Moon's Face: A Study of the Origin of its Features. Presidential address, Phil. Soc. of Washington, 1892, Bull. Vol. XII, pp. 241–292.

VOLCANOES.

Number of volcanoes.—It is impracticable to state exactly the number of volcanoes that are active at the present time, because most volcanoes are periodic, and become active at more or less distant periods, and it is impossible to say whether a given volcano that may be now quiescent has really become extinct or is only enjoying its customary period of rest. It is quite safe to include at least 300 in the active list, and the number may reach 350 or more. The numbers that have been active so recently that their cones have not been entirely worn away is several times as great.

Distribution of Volcanoes.

1. **In time.**—In the earliest known ages igneous action appears to have been very general, if not practically universal. No area of the earliest (Archean) rocks is now known which is not formed chiefly of rocks that appear to have been either intruded or extruded. Rocks which can reasonably be assigned to the hypothetical molten globe, if there be such, are not here included. It is probable that the surface of the early earth was as thickly occupied with points of extrusion as the surface of the moon appears to be. In the ages between the Archean and the present, the distribution of volcanic action over the surface seems to have been in a general way much what it is to-day; that is, certain areas were volcanically active at times, while other and larger areas were measurably free from any outward expressions of igneous action. This is not equally true of all ages, as will be seen in the historical studies that follow. There were periods when volcanic activity seems to have been widespread and energetic, and others when it was limited both in amount and distribution. The known facts do not indicate a steady decline in volcanic activity, but rather a periodicity; at least this is so for the portion of the globe that is now well enough known geologically to warrant conclusions. One of the greatest of the volcanic periods falls within the Cenozoic era, just preceding the present geological period, and the volcanic activity of the present is perhaps but a declining phase of that time.

2. **Relative to land and sea.**—At present the active volcanoes are chiefly distributed about the borders of the continents, and, less notably, within the great oceanic basins. On this account the sea has often been supposed to have some connection with volcanic action, and the

presence of chlorine in the volcanic emanations has been cited in support of this position. When critically examined, however, the argument from distribution is not very strong; for the volcanoes are not distributed equally or proportionately about the several oceans, as if dependent on them. Volcanoes are especially numerous around and within the Pacific, the greatest of the oceans, and this might seem a favorable instance, but they are also numerous around and within the Mediterranean, a relatively small body of water. Volcanoes are not especially abundant in or about the margins of the Atlantic.

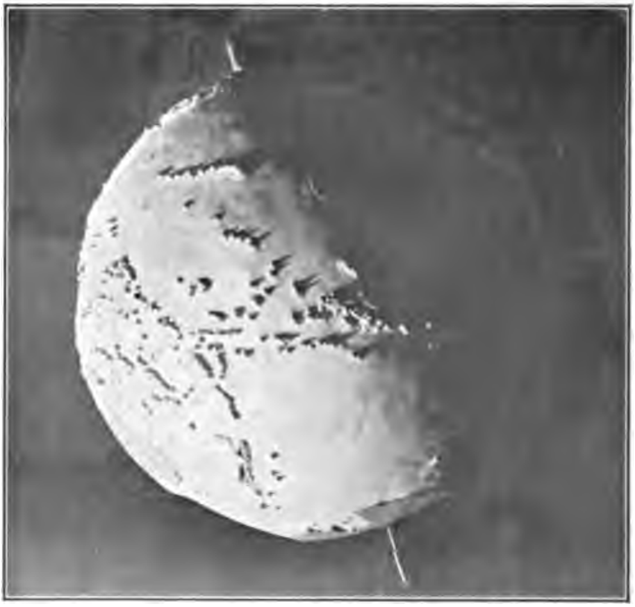

Fig. 460.—Volcanoes in the Pacific. Jones Relief Globe. (Photo. by R. T. Chamberlin.)

If volcanoes were dependent upon proximity to the sea, the relation should be close in the past as well as in the present, but this does not seem to be true. There has recently been much volcanic activity in the plateau region of western America at long distances from the Pacific basin. Even on the plains east of the Rocky Mountains notable volcanic action took place. There were also volcanoes in the interior of Asia and of Africa.

3. **Relative to crustal deformations.**—The distribution of present and recent volcanoes is much more suggestively associated with those portions of the crust *that have undergone notable changes* in position in comparatively recent times. The great "world-ridge" stretching from Cape Horn to Alaska and thence onwards along the east coast of Asia is a striking instance, for it is dotted throughout with active and recently extinct volcanoes. The tortuous zone of mountainous wrinkles that borders the Mediterranean and stretches thence eastward to the Polynesian Islands is another notable volcanic tract. These two belts include the greater number of existing and recent volcanoes on the land, while the great basins associated with them embrace the chief oceanic volcanoes.

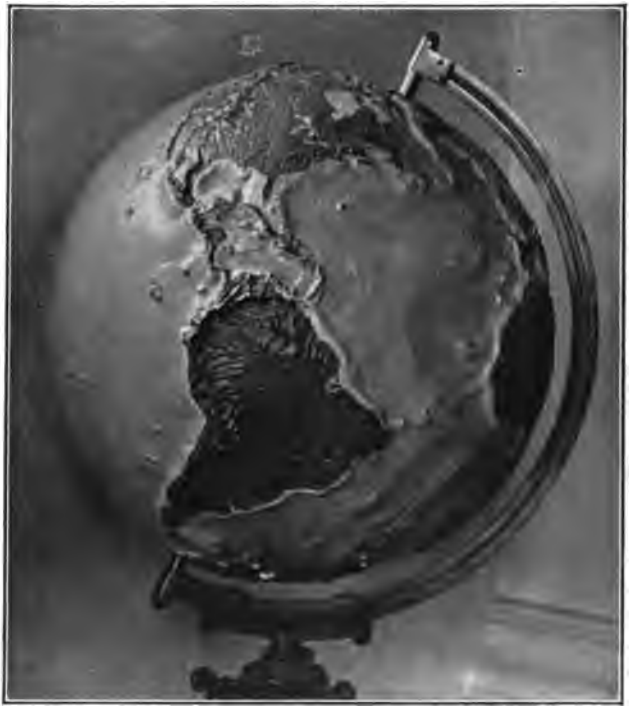

FIG. 461.—Active volcanic area at the junction of the continental segments of North and South America, and of the abysmal segments of the Atlantic and Pacific. Jones Relief Globe. (Photo. by R. T. Chamberlin.)

There is perhaps some significance in the fact that the most active regions of vulcanism to-day lie *at the angular junctions of the great earth-*

segments. The Antillean and Central American volcanic region, that has recently been so demonstrative, lies where the southern angle of the North American continental block joins the northern angle of the South American continental block, and where the western angle of the North Atlantic abysmal segment closely approaches one of the eastern angles of the great Pacific abysmal segment. The complex and very active Java-Philippine volcanic region lies where the southeastern angle of the great Asian segment projects toward the Australian block, and where the western angle of the Pacific block approaches the northeastern angle of the Indian oceanic segment. The active Alaskan volcanic area lies at the angles of the North American, Asian, Pacific, and Arctic segments. The Mediterranean volcanic area falls less notably under this generalization, but it lies where the continental blocks of Europe and Africa come into peculiar relations to each other on either side of the remarkable Mediterranean trough. The eastern angle of the North Atlantic segment is near by, but not in very close relations. The Icelandic region, small but vigorous, lies near the junc-

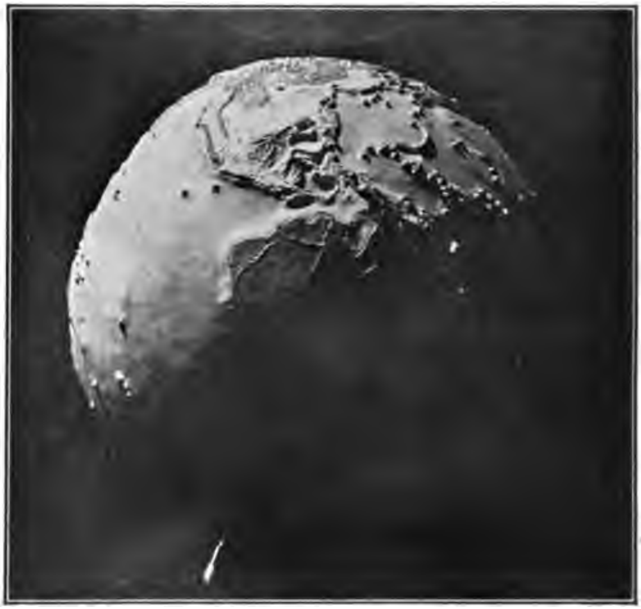

FIG. 462.—Active volcanic area at the junction of the continental segments of Asia and Australia, and the abysmal segments of the Pacific and Indian oceans. Jones Relief Globe. (Photo. by R. T. Chamberlin.)

tion of the North American, European, North Atlantic, and Arctic segments, and the New Zealand volcanic region is somewhat less closely related to the approach of the Australian, Antarctic, Pacific, and Southern oceanic segments. Nearly all of these angular conjunctions involve two depressed segments joining two relatively elevated segments. This relationship suggests a causal connection between the intensified movements at these angular conjunctions and the intensified volcanic action of these regions. There are enough volcanoes, however, that do not fall into these groups, or apparently into any other grouping, to suggest that the development of volcanoes is not wholly dependent on any surface relationship, but that it is connected with deep-seated causes that are indeed modified, but not wholly controlled, by surface conditions, or even by the movements of the master segments of the earth's crust.

4. **In latitude.**—The distribution of volcanoes appears to have no specific relation to latitude. Mounts Erebus and Terror, amid the ice-mantle of Antarctica, and Mount Hecla in Iceland, as well as the numerous volcanoes of the Aleutian chain, give no ground for supposing that volcanoes shun the frigid zones. On the other hand, the numerous volcanoes of the equatorial zone do not imply that they avoid the torrid belt. Their distribution appears to be independent of latitude. This is not cited because of any supposed effects of external temperature, for that must be trivial, but because it bears on the question whether strains are now arising from the supposed *slackening of the earth's rotation,* which have any connection with volcanic action. If the oblateness of the earth is decreasing, the equatorial belt must be sinking and growing shorter, and hence must be under lateral pressure, while the polar caps must be rising, and increasing their curvatures, and should be under tension. These conditions, if real, might be supposed to have something to do with the extrusion of lava. Nothing in the present or the past distribution of igneous action seems to afford much support to this hypothetical inference.

5. **In curved lines.**—In the Antilles, the Aleutian Islands, the Kurile Islands, and in other instances, there is a notable linear arrangement of volcanoes with appreciable curvature. It has been noted that the convexity of the curves is turned toward the adjacent ocean. In some cases, however, there is a notable linear arrangement without appreciable curvature, as in the Hawaiian range, in the recently extinct line

of cones of the Cascade Range, and in others. Less often, volcanoes are bunched irregularly, as in some of the groups of volcanic islands of the Pacific (Fig. 460).

Relations of Volcanoes.

1. **Relations to rising and sinking surfaces.**—So far as observations cover this point, the area immediately adjacent to active volcanoes is rising (Dutton). This is shown by raised beaches, terraces, coral deposits, etc. Whether this is wholly due to the expansional effect of the heating of the subterrane by the rising lava, or whether it has a wider significance, is not known. If a broader view is taken, it does not appear that there are sufficient data to connect volcanic action exclusively with either the rising or the sinking of the general surface. It is certain that the great mountain ranges and plateaus in which so much of the more recent volcanic action has taken place have been recently elevated relatively, but they have also undergone more or less of oscillation, involving some relative depression. The question whether the Pacific basin as a whole has been relatively elevated or depressed in modern times is a mooted one. Darwin[1] and Dana,[2] as the result of their early studies on its coral deposits and on other phenomena, concluded that the Pacific was a sinking area, but this view has been recently challenged by Murray [3] and Agassiz [4] with at least some measure of success. From the fiords on the borders of the Pacific and other physical phenomena, the inference has been drawn that relative sinking of the land has recently taken place. Raised beaches on the coasts are interpreted as indicating a relative rise of the land or a sinking of some ocean basin, for the withdrawal of the waters can only be the result of increasing the capacity of the oceanic basin as a whole. The most probable view is that the general areas of present and recent volcanic action are partly rising areas and partly sinking areas, and that movement of either kind may be connected with the extrusion of the lavas. The rising and sinking are but complementary phases of a deformation of the earth's body, and involve a readjustment of stresses within the

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[1] Structure and Distribution of Coral Islands.

[2] Corals and Coral Islands.

[3] Proc. Roy. Soc. Edin., Vol. X, pp. 505–18, and Vol. XVII, pp. 79–109; Nature, Vol. XXXII, p. 613; Narrative Chal. Exp., Vol. I, pp. 781–2.

[4] Bull. Mus. Comp. Zool., Vol. XVII, 1889.

body of the earth. These stresses are possibly an essential factor in eruptions.

2. **Relations to one another.**—A most significant feature of volcanic action is the degree of concurrence or of independence of action in adjacent volcanoes. In some instances they act as though in sympathy, as in the recent outburst in Martinique and Saint Vincent, and the concurrent symptoms of activity in other places. On the other hand, the independence of neighboring vents is sometimes extraordinary. The group of volcanoes near the center of the Mediterranean, of which Vesuvius and Etna are the most conspicuous examples, usually act with measurable independence of one another, an eruption in the one not being habitually coincident with an eruption in the others. But the most conspicuous instance of independence is found in the great craters of Mauna Loa and Kilauea in Hawaii. They are only about twenty miles apart, the one on the top and the other on the side of the same great mountain mass. The crater of Mauna Loa is about 10,000 feet higher than the crater of Kilauea, and yet, while the latter has been in constant activity as far back as its history is known, the former is periodic. The case is the more remarkable because of the greatness of the ejections. The outflow of Mauna Loa in 1885 formed a stream from three to ten miles in width, and forty-five miles in length, with a probable average thickness of 100 feet, and some of its other outflows were of nearly equal greatness; indeed its outflows are among the most massive that have issued from volcanoes in recent times. Besides this massiveness there have been extraordinary movements of the lava within the crater, if the testimony of witnesses may be trusted. But throughout these great movements in the higher crater, the lava-column of Kilauea, 10,000 feet lower, continued its quiet action without sensible effects from its boisterous neighbor. The bearing of such extraordinary independence upon the sources of volcanic action is very cogent, for the lavas are of the same type, both being basalts, that of Mauna Loa being notably basic and probably as high in specific gravity as that in Kilauea. No difference in specific gravity that could at all account for a difference in height of 10,000 feet can be presumed, unless their ducts remain separate to extraordinary depths. Nor does it appear possible that a superior amount of gas within the column of Mauna Loa could account for such an extraordinary difference in height, for the hydrostatic pressure of such a column is not far from 10,000 pounds to the square inch.

Even if the difference in the heights of the columns could be explained by differences in specific gravity, the agitation of the one should be communicated to the other, and an outflow of the one, particularly an outflow by a breakage through its walls sufficient to lower its surface hundreds of feet, as has repeatedly occurred in Kilauea, should change the surface of the other proportionately, if they were in hydrostatic equilibrium. It seems a necessary inference, therefore, that the two lava-columns have no connection with each other or with a common reservoir. The tops of some lava-columns stand about 20,000 feet above the sea, while others emerge on the sea-bottom far below sea-level. The total vertical range is, therefore, probably between 30,000 and 40,000 feet, a difference which tells its own story as to their relative independence.

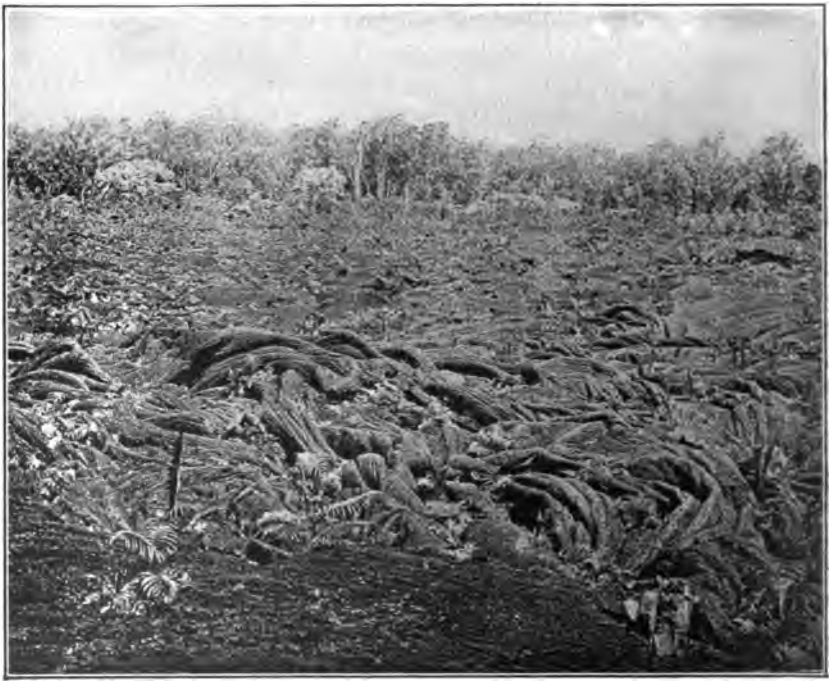

Fig. 463.—Surface of lava-flow of 1881, from Mauna Loa, as seen back of Hilo, Hawaii. (Photo. by Calvin.)

3. **Unimportant coincidences.**—Eruptions seem to be somewhat more liable to occur at times of high atmospheric pressure than at low, doubt-

less because the increased atmospheric weight on a large area of the adjacent crust aids in forcing out the lava or the volcanic gases. This can only be effective when other forces have almost accomplished the result, and would doubtless have completed it a little later had not the atmospheric wave supplied the little remaining pressure needed. Eruption seems also to be more common when the tidal strains favor it, for like reasons. In the same class are probably to be put the effects of heavy rains, whether they act by gravity or by giving rise to steam. Such agencies are to be regarded as mere incidents of no moment in the real causation of vulcanism, but of some value in determining the precise moment of action. This is not to be understood as inconsistent with the view that the periodic stresses of the body-tides of the earth are important factors in vulcanism, as elsewhere explained, but merely that the special time of surface-eruption is only incidentally connected with the water-tides.

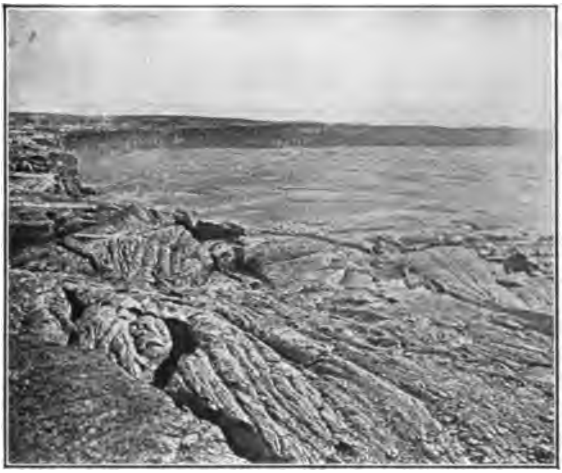

Fig. 464.—Crater of Kilauea.

Periodicity.—Most volcanoes are periodic in their stages of action. Long dormant periods intervene between eruptive periods. Volcanoes supposed to be extinct occasionally awaken with terrific violence. Sometimes also they awaken quietly. This larger periodicity yet awaits an explanation, but it very likely means a temporary exhaustion of the supply of gas or of lava, or of both, to which the active stage is due.

Formation of Cones.

Lava-cones.—The lava usually flows away from the vent in short streams which solidify before running far. As the lava-streams flow in different directions at different times, the total effect is a low cone formed of radiating tongues surrounding the point of exit. Occasionally the streams run a dozen or a score of miles, but such cases, except in the gigantic volcanoes of Hawaii and a few others, are rare. Often the streams congeal before they reach much beyond the base of the cone, and quite often while they are yet on its slope. So far, therefore, as the volcanic cone is formed of lava, it has a radiate structure made up of a succession of congealed lava-streams. In these cases the slopes are low, because the fluidity of the lava prevents the development of high gradients. It is, however, rather the exception than the rule, that the cone is made up mainly of lava-streams, though the great Hawaiian volcanoes are of this class.

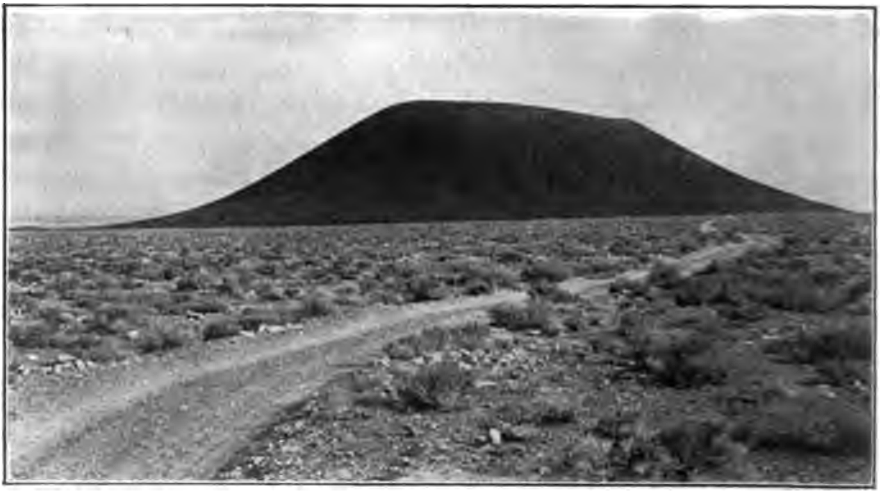

FIG. 465.—Typical cinder-cone, Clayton valley, Cal. (Turner, U. S. Geol. Surv.)

Cinder-cones.—The larger portion of the lava blown into the air by the expanding gas-bubbles falls back in the immediate vicinity of the vent and builds up a cinder-cone. From the nature of the case, this often takes on a beautiful symmetry and assumes a steep slope (Fig. 465). The ragged cinders lend themselves readily to the formation of an acute cone, quite different from the flatter cone formed by lavas. Sometimes

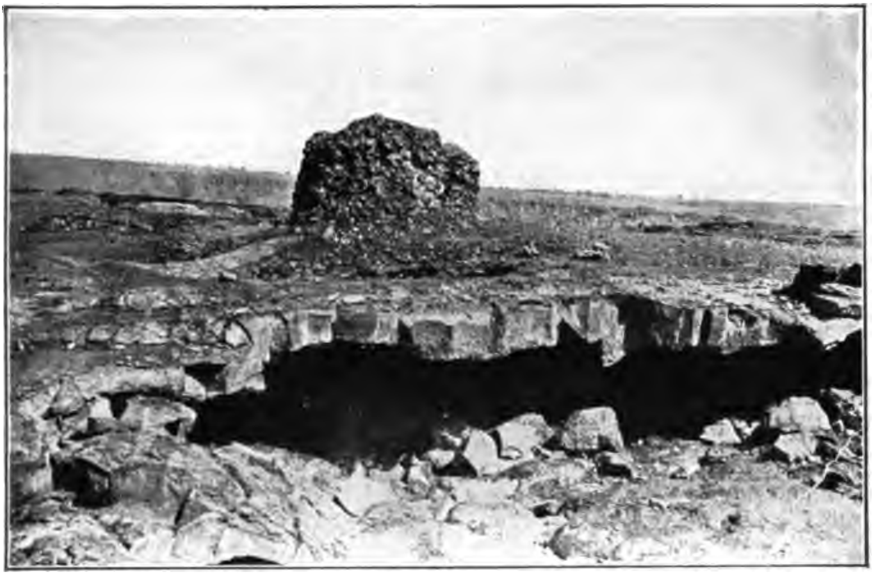

FIG. 466.—Spatter-cone and cavern. Kilauea, Hawaii. (Photo. by Libbey.)

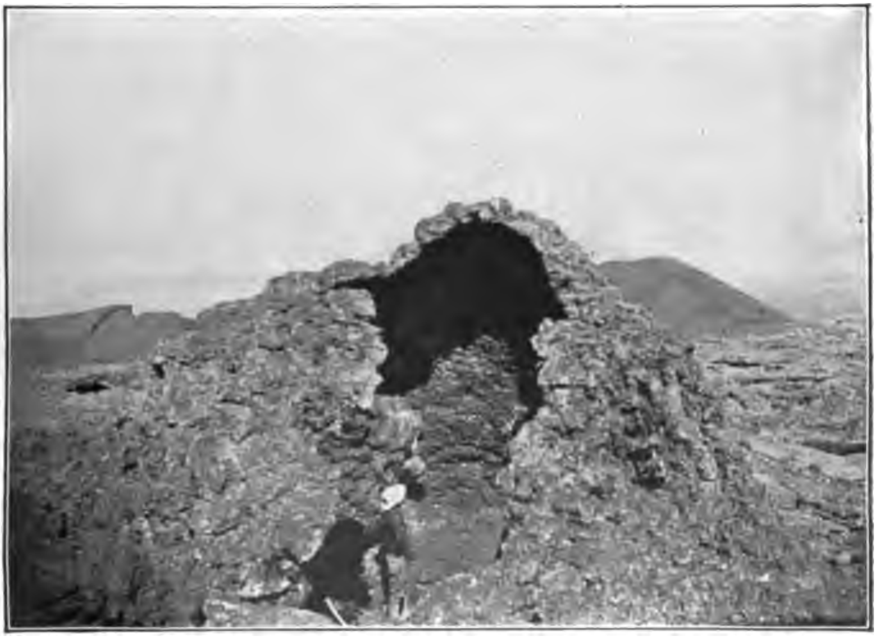

FIG. 467.—Hollow spatter-cone. Oregon. (Russell, U. S. Geol. Surv.)

the cinders are still plastic when they fall, and weld themselves together and hold their places even on very steep slopes, but usually they have already hardened before they reach the surface.

Sudordinate cones.—Small or temporary vents formed as offshoots from the main vents often give rise to secondary or "parasitic" cones. These are sometimes numerous, as in the case of Etna, and they may be so important that the mountain becomes a compound cone. A still more subordinate variety consists of "spatter-cones" formed by small mildly explosive vents that spatter forth little dabs of lava which form chimneys, or cones, and sometimes completely curved domes over vents (Figs. 466 and 467). Spatter-cones often arise from the lava-flows themselves.

Composite cones.—From most existing volcanoes there issue both lava-flows and fragmental ejecta, and the resulting cones are composite in material. The lava more frequently breaks through the side of the cone than overflows its summit, and this gives rise to irregularities of form and structure. The cones are also subject to partial destruction both by the outbursts of lava and by the explosions, and perhaps also by migration of the vents. As a result, many volcanic regions show old, partially destroyed craters, together with new and more perfect ones, and the history of volcanic action in a region may often be read in the succession of cone formations.

The form of the cone, when composed chiefly of lava, is also affected by the mass of the outflow and by its fluidity. The larger the outflow at a given time, other things being equal, the wider it distributes itself and the flatter is the cone. As a rule, the basic lavas are more fluid than the acidic, and the cones of basic lavas are flatter than the cones of acidic lavas.

Extra-cone distribution.—In violent eruptions, the steam, accompanied with much ash, is shot up to great heights, often rolling outwards in cumulus or cauliflower-like forms (Fig. 458). In the more violent explosions these columns are projected several miles. In the phenomenal case of Krakatoa the projection was estimated at seventeen miles. The steam, by reason of its great expansion and its contact with the colder regions of the upper air, is quickly condensed, and prodigious floods of rain frequently accompany the eruption. This rain, carrying down a portion of the ash and gathering up much that had previously fallen, gives rise to *mud-flows*, which in some cases constitute a large part of the final deposit. These mud-flows chiefly lodge on the lower slopes of the volcano or adjacent to its base, and give rise to rather flat cones, sometimes designated as *tufa-cones* to distinguish them from cinder-cones formed by the direct fall of fragmental material. Mud-flows appear also to be formed by the ejection of mud and water that had gathered in quiescent craters during intervals between stages of eruption.

A portion of the finer exploded material floats away in the air to greater or less distances, and forms widespread *tufa-deposits*. In some cases beds of volcanic ash of appreciable thickness (as those of Nebraska)[1] are found far from any known volcanic center. The extremely fine ash from the great explosion of Krakatoa floated several times around the earth in the equatorial belt and spread northward into the temperate zones.

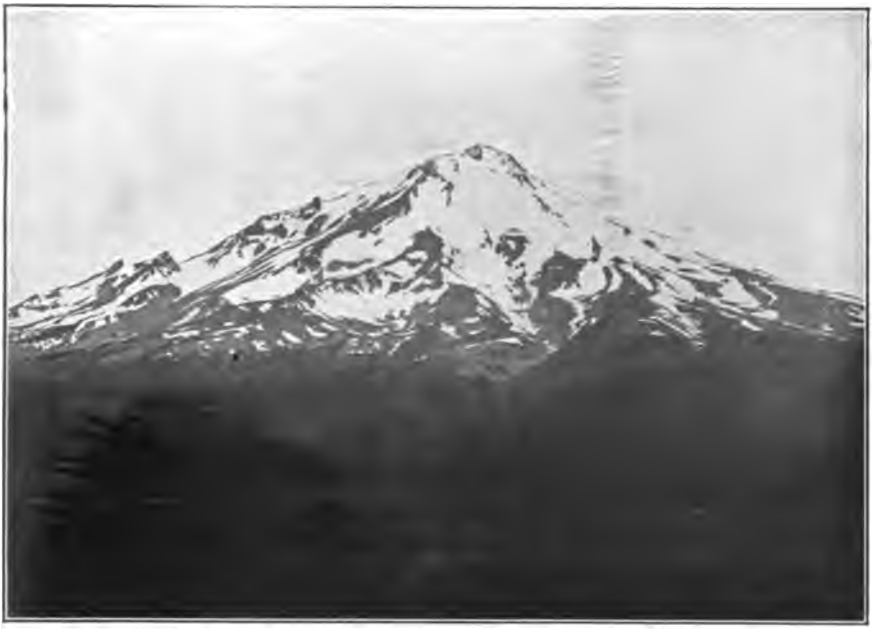

FIG. 468.—Mt. Shasta, a typical extinct cone, furrowed by erosion, but retaining its general form. (Diller, U. S. Geol. Surv.)

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[1] Ante, p. 22.

LAVAS.

Their nature.—In the chapter on the Origin and Descent of Rocks, the nature of lavas and of the rocks derived from them has been discussed (Chapter VII). In view of prevalent misconceptions, it may be repeated, for the sake of emphasis, that lavas are mutual *solutions* of mineral matter in mineral matter, rather than simply melted rock. Into this mutual solution there enter not only rock materials, but gases. The distinction between mutual solutions and simple molten rock cannot be rigorously made, but it is at least essential to know that the minerals do not necessarily crystallize from lavas in the order of their melting temperatures, or in any uniform order, but rather in the order in which saturation of the several mineral constituents happens to be reached in the given mutual solution. Thus quartz, which has a very high melting-point, is often one of the last minerals to crystallize. The mutual solutions are exceedingly complex, embracing a wide range of chemical substances, but the chief of them, as already stated, are silicates of aluminum, potassium, sodium, calcium, magnesium, and iron, with minor ingredients of nearly all known substances, in greater or less proportion. The old idea of lavas as simply melted rock is not, however, wholly to be abandoned. The mode of solidifying is often

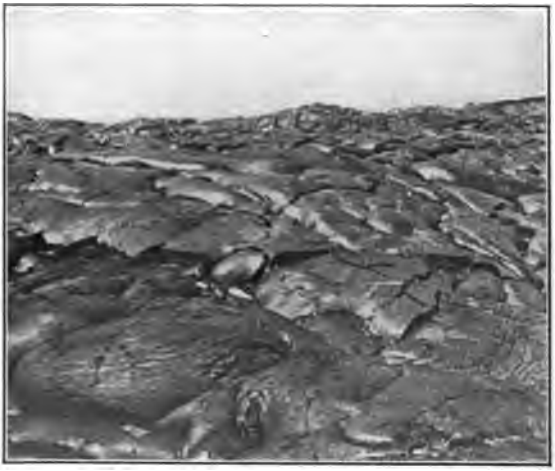

Fig. 469.—Lobular form of lava-flow, "Pahoehoe." (Dutton, U. S. Geol. Surv.)

simply that of molten matter freezing. If lava be suddenly cooled, the congelation is essentially the solidification of a melted substance. The result is a glassy body, every part of which has essentially the same composition that the liquid had. Usually, however, even in this case, the gases escape in part. If the cooling is slower, the various substances in the mixture crystallize out into minerals in the order in which they severally reach saturation. This involves the principle that solubility is dependent on temperature, and that as the temperature sinks the degree of solubility declines, and the saturation-point for some constituents of the solution is reached earlier than for others. With sufficiently slow cooling, all the material will pass into the solid state by the crystallizing of the several minerals in succession. This does not mean that two or more minerals may not be forming at the same time, for crystals often interfere with each other's growth. It does, however, involve the doctrine that some substances may complete their

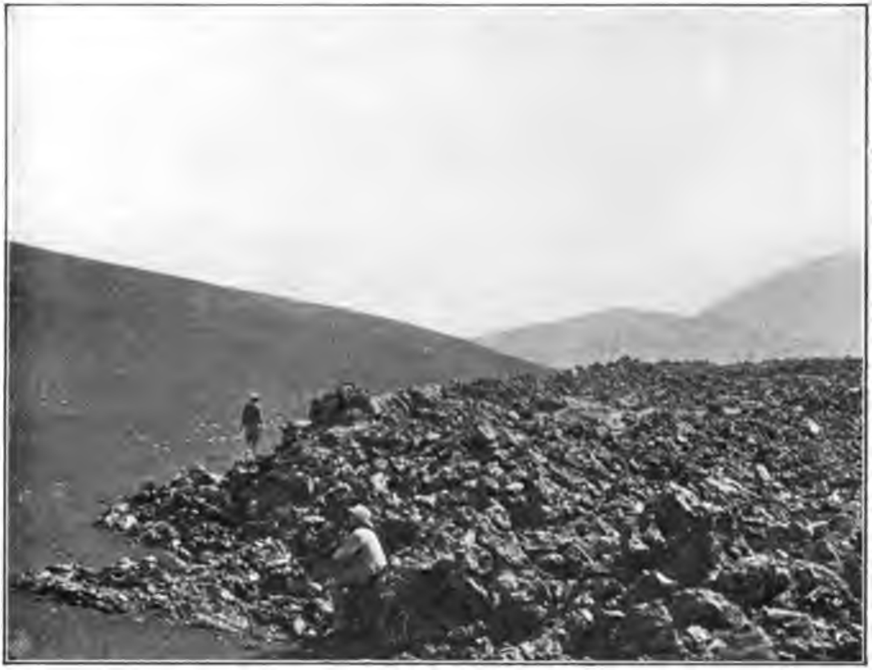

FIG. 470.—Terminal portion of a rough lava-flow, "*aa*." Cinder Buttes, Idaho. (Russell, U. S. Geol. Surv.)

crystallization while the surrounding material is yet in the fluid condition. In most igneous rocks nearly perfect crystals of certain minerals are common, while other minerals, crystallizing later, are compelled to adapt themselves to the space left. This conception is supported by the fact that lavas, while still in the fluid condition, often contain well-formed crystals, and these crystals sometimes make up a considerable percent. of the flowing mass, just as water in certain conditions may be filled with crystals of ice. So also crystals after having been formed may be redissolved in part, doubtless because of changes in the nature of the magma due to undetermined conditions which may arise in the process of crystallization, or from the accession of gas, or from new material dissolved from the walls of the passageway.

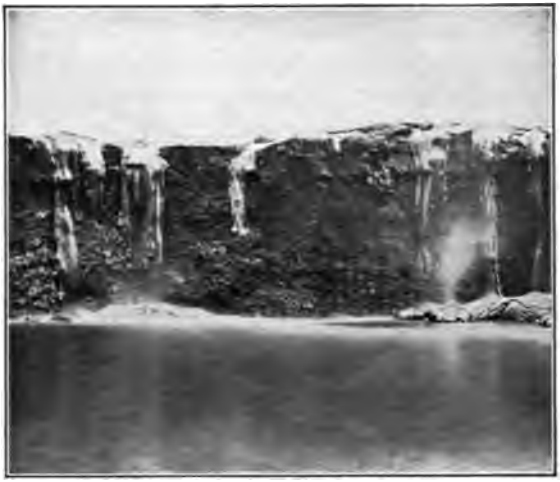

FIG. 471.—Lava flowing over a precipice near Hilo, Hawaiian Islands.

Consanguinity and succession of lavas.—The lavas that are poured forth at different stages in the succession of eruptions of a given region are usually not the same, as might naturally be expected, but form a curious series the members of which are related to one another. Iddings has called this relation *consanguinity*.[1] No universal law of succession has yet been established, and perhaps none exists; but Richthofen[2] many years ago announced a definite order for the Tertiary flows of western America which seems to hold fairly well in its general aspects, though not everywhere completely realized, so far as surface observation goes. Richthofen's order is: (1) lavas of neutral types, (2) lavas of acid types, (3) lavas of basic types, (4) lavas of more acid types, and (5) lavas of more basic types. The special varieties of rock vary, and even the general order is often apparently defective. The defects are sometimes assigned to the concealment of some of the outflows. While this may be true in some cases, it is not unlikely that in others there is a real failure of the sequence. At any rate, the sequence can only be regarded as a rough generalization. It is supposed to be due to magmatic differentiation caused by the differences of temperature to which the different parts are subjected underground, by differences of specific gravity and fluidity which result from changes of temperature, and probably by other causes.

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[1] Origin of Igneous Rocks. Phil. Soc. of Wash., Vol. XII, pp. 89–214.

[2] The Natural System of Volcanic Rocks. Cal. Acad. of Sci., 1868.

Temperatures of lavas. — Accurate determinations of the temperatures in the center of the lava-columns, where they have been least reduced by contact with the rock-walls, have not yet been made, but it is clear from the whiteness of the lavas that their temperatures are often appreciably above the melting-point. This is also a necessary inference from the length of time they remain fluid, notwithstanding the great surface contact of the column in its miles of ascent, the conversion of contact water into steam, and the expansion and escape of the gases. In cases where determination has been practicable (and they certainly do not represent the maximum temperatures) it has been found that the melting-points of silver, about 960° C., and of copper, about 1060° C., are reached. In connection with overflows, it has been found that brass is decomposed into its component metals, the copper actually crystallizing. Silver has been sublimed, and made to redeposit itself in crystalline form. This implies much more than the bare melting temperatures. Even the fine edges of flints have been fused. It is, therefore, probably safe to assume that the original temperatures of the lavas as they rise to the surface sometimes reach considerably beyond 2000° Fahr. (1093° C.), and may perhaps even attain 3000° Fahr. or more. Even these temperatures must be somewhat below the original subterranean temperatures of the lavas, because some heat must necessarily be lost in rising, partly by contact with the walls of the colder rocks through which they pass, probably for as much as a score of miles at least, and partly from the expansion of the gases within them. If any considerable part of these gases is derived from waters which joined the lava in its upward course in the fracture zone, the energy consumed in raising the water to the high temperatures of the lavas must be subtracted from the original heat, and must be a further source of reduction of temperature. It is important to emphasize this point in view of its bearing upon the origin of the lavas. It has been suggested that lavas may be due to an aqueo-igneous fusion, a kind of fusion which may take place at comparatively moderate temperatures. It seems obvious, however, from the phenomena themselves, that temperatures as high as ordinary dry fusion, and perhaps even higher, are attained. It is clear also that the maintenance of the liquid condition in a constant state of ebullition for a long period of time implies a large surplus of heat above that necessary for liquefaction simply. This is especially true if the ebullition comes from surface-waters penetrating to and becoming absorbed in the lava-column below. This process must tend rapidly to exhaust the heat in the column of lava. If, on the other hand, the gases are derived from the deep interior, and the ebullition at the surface is due to their escape, they may bring up new supplies of heat to counteract the cooling effects of their expansion.

Depth of source.—Attempts have been made to ascertain the depth from which lavas rise, by means of the earthquake tremors that accompany eruptions. The estimates have ranged from seven or eight to thirty miles. The mode of estimate is that discussed under earthquakes, and is subject to the corrections there indicated. If these could be perfectly applied, the estimates might probably all fall within ten miles, and not improbably all within six miles of the surface. But in any case the method really tells very little as to the true point of origin of the lava. At most it probably only tells where the ascending lava begins to *rupture* the rock through which it passes, and rupture may not be possible below the zone of fracture, which is probably not more than six miles deep. In the zone of flowage below, where the pressure is too great to permit fracture, the lava not improbably makes its way by some boring or fluxing process, which might not, because of its nature, be capable of giving rise to seismic tremors. The behavior of the tremors perhaps forces us to locate the origin of lava movement *at least as low as the bottom of the fracture zone*, but it probably offers no sufficient ground for limiting the lava's origin to this or any other specific depth.

VOLCANIC GASES.

The most distinctive feature of volcanoes is the explosive action arising from the gases and vapors pent up in the lava. There is not a little explosive action of a secondary character arising from the mere outer contact of surface-waters with lavas or with the hot rocks of the crater walls, or with the hot ashes and rocks thrown out; but these are incidental, not essential, features.

The precise nature of the occlusion or absorption of gases and vapors has not yet been determined. It is thought that lava spontaneously absorbs such gases when at high temperatures, and especially when the gases are under great pressure, and that as the pressure is relieved and the lava is cooled and solidified, the larger part of the gases escapes. In those cases in which the eruption is quiet, the escape of the gases is but partial while the lava is in the crater, and much gas remains to be given out from the molten material after it has been extruded and is about to congeal. The gases are then given off with relative slowness and quietness. If, however, the lavas are surcharged with gases, and if these are restrained from free escape by the viscosity of the lavas, the gases gather in large vesicles in the lava in the throat of the volcano, and on coming to the surface explode, hurling the enveloping lava upwards and outwards, often to great distances. The violence of projection reduces a portion of the lava to a finely divided state constituting the "ash" and "smoke" of the volcano. Other portions less divided are inflated by the gases disseminated through them, and form "pumice" and "scoria," according to the degree of inflation, while masses of lava that have already solidified into more or less rounded masses in the crater are hurled forth as "bombs"; not infrequently portions of the walls of the crater or of the duct below are also disrupted and shot forth.

Differences in gas action.—The causes of the differences of gas action in different volcanoes are undetermined, but the following suggestions may point to a part of the truth: (1) Doubtless some lavas contain more gases than others, and hence are predisposed to be more explosive; (2) some are more viscous than others and hence hold the gases more tenaciously until they accumulate and acquire explosive force, while the more liquid lavas allow their gases to escape more freely and easily; (3) some are hotter than others, and hence hold their gases until after they have escaped from the crater, when they give them off from their expanded surfaces in the open air, where there is no restraint to develop explosiveness; (4) some flows are so massive that they cool to the chief gas-discharging point only after they are spread out on the surface, when quiet escape is possible; (5) probably a main occasion of the very violent explosions lies in the fact that the lavas have begun to crystallize while yet in the duct of the volcano. The crystals, in forming in the magma, exclude the gases from themselves, and this excluded portion overcharges the remaining portion of the lava. This process continues as the lava rises and grows cooler until the gases acquire great volume and explosive force. This view is sustained by the fact that the pumice and ash of such extraordinarily explosive eruptions as those of Krakatoa and Pelée contain many small crystals which had certainly formed before the explosive inflation took place. Incipient crystallization does not, however, appear to be a universal accompaniment of explosive action.

Spasmodic action.—The discharge of the gases is spasmodic, and usually consists of a succession of distinct explosions. Sometimes these succeed one another at rather constant and frequent intervals, as in Stromboli, where the explosions follow one another at intervals of three to ten or more minutes. In many others the outbursts are rhythmic, while in others the spasms are distant and irregular.

Kinds of gases.—Steam is the chief volcanic gas. Its constituents, hydrogen and oxygen, are also present in the free state, and are perhaps the result of the dissociation of the steam at the very high temperatures of the lavas. Carbon dioxide is probably next in abundance. No positive statement as to the relative amounts of the subordinate gases can be made because of the obvious difficulties of obtaining anything like a representative analysis of the gases concerned in the great volcanic eruptions. The materials for the analyses which have been made were derived chiefly from little secondary or "parasitic" vents, or from side-wall crevices, through which the volcanic gases rise. Such vents probably derive their gases from the very border of the main mass, where it is most subject to the influence of waters and gases from the adjacent walls, and it is uncertain how far they are truly representative of the gases in the interior of the lava itself. The data now at command seem to indicate that carbon dioxide increases greatly in relative abundance as volcanic action dies away. Great quantities of this gas are often given forth long after all signs of active vulcanism have disappeared. Such gases have been attributed to the action of the lavas on buried beds of limestone or other carbonates, but in many cases the geology of the region offers no special support to this hypothesis. It does not seem inherently probable that the heat of the lava would be sufficient to decompose limestone at a period very long after the active eruption. An alternative suggestion is that the stronger volcanic acids mentioned below are gradually conveyed into the adjacent rocks and there act on limestones or on partially carbonated crystalline rocks, setting free carbon dioxide. Whatever may be true with regard to secondary gases of this kind, it is quite certain that the lavas themselves contain large quantities of carbon dioxide, and also of carbon monoxide, doubtless reduced from the dioxide. Sulphur gases are very common accompaniments of volcanic eruptions. They take the forms of sulphuretted hydrogen and sulphurous acid and perhaps of sublimated sulphur, all of which are liable to pass by oxidation and hydration into sulphuric acid. Chlorine and hydrochloric gases are also common, particularly at high temperatures. Fluorine and other gases are occasionally present. Certain gases, such as hydrogen and chlorine, are especially associated with high temperatures and energetic action, and are probably dependent on them. Hydrochloric acid and the sulphurous gases are also mainly associated with high temperatures, while sulphuretted hydrogen is commoner at lower temperatures. Oxygen, nitrogen, and probably carbon dioxide or carbon monoxide are present throughout all ranges of temperature. Nitrogen is a rather frequent but not very abundant constituent of the volcanic gases. How far it results from admixture of the atmosphere and how far it is original, is not determined. It is, however, one of the gases found in volcanic rocks after they have cooled, and is presumably original in part. A large series of secondary vapors naturally arise from the volatilization of substances contained in the lavas, such as the oxides, chlorides, and sulphides of the metals, etc.

Residual gases in volcanic rocks.—Some light upon the vital question of the original, as distinguished from the secondary gases of lavas may be found in the analyses of the gases that remain in the lavas after they are solidified. When the lavas lodged underground without free communication with the surface, there is reason to think that they retained a larger percentage of their original gases in solidification than in cases of free exposure at the surface; at any rate, such rocks contain notable quantities of gases occluded in some way within themselves. Recent surface-lavas also contain gases of similar kinds, but not in equal degree, so far as available analyses show. The gases are in part held in numerous small cavities within the constituent minerals, especially in the quartz. This is perhaps due to the fact that quartz usually crystallizes late in the process of solidification, and its mother-material becomes crowded with gases excluded by the previous crystallization of other minerals. Analyses of twenty-five crystalline rocks of various kinds from many typical localities by Tilden,[1] gave an average volume of gas, under ordinary atmospheric pressure, four and a half times that of the containing rock. This shows the condensed condition in which the gases are held. Of these gases, the chief is hydrogen, which much exceeds all the rest. Next in order of abundance is carbon dioxide, followed by carbon monoxide, marsh gas (CH_4), and nitrogen. Water is frequently present and free oxygen almost universally absent. The average ratio of hydrogen to carbon dioxide by volume in these analyses is about 70 : 30. Five complete analyses gave the following averages: H_2, 52.134; CO_2, 34.104; CO, 8.422; CH_4, 3.224; N_2, 2.072. It will be seen that the gases contained in these rocks are in proportions radically different from those of the atmosphere, and it is doubtful whether they can be reasonably assigned to any other source than the lavas from which the rocks were formed. It is to be noted, however, that some sedimentary and meta-sedimentary rocks, such as quartzite and quartz-schist, contain similar gases, but this may be because the granules of the original rock retain them, notwithstanding the secondary processes through which they have passed. Analyses of meteorites show essentially the same gases in much the same proportions. If evidence of this kind can be trusted, the standard original gases of lavas are the elements or compounds of hydrogen, carbon, and nitrogen, in the order named, while the chlorine and sulphur gases are to be regarded as accessory. Because of their intensely energetic and noxious character, these latter gases make themselves disproportionately manifest in the vicinity of active volcanoes. That they are really not preponderant seems to be implied by the fact that the volcanic rains, which are extremely copious, are usually fresh, and only in rare cases is the presence of the hydrochloric or sulphurous elements sufficient to produce noxious effects. Volcanic and meteoric data seem to indicate that steam is held less tenaciously than the other gases in the magmas as they solidify into rocks.

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[1] Chemical News, April 9, 1897.

The source of the gases.—As already noted, it is one of the outstanding problems of geology to determine (*a*) how far the gases of lavas were possessed by them from their origin, whatever that may be, and (*b*) how far they have been acquired in the lava's ascent to the surface. It is recognized that lavas have the power of absorbing gases, and one of the views entertained is that surface-waters, percolating through the rocks and coming in contact with the ascending column of lava, are converted into steam, which is absorbed into the lava and rises with it to the surface. There are two phases of this view. (1) The more conservative one supposes that the water merely penetrates the fracture zone of the surface of the earth through the ordinary means of passage of underground-waters, and so makes a comparatively short circuit. Under this view, the steam and other gases given forth are not a contribution to the atmosphere and hydrosphere, but merely a restoration to them of water and dissolved gases previously carried down from the surface. An even narrower view is sometimes entertained which supposes that the larger part of the water descends through the volcanic cone itself, or immediately about its base. The presence of chlorine gases in the volcanic emanations and the nearness of most existing volcanoes to the sea have been the basis for the idea that sea-water, penetrating to the lava, is a chief source of the volcanic gases. (2) The broader phase of the view assumes that the waters penetrate not only the outer fracture zone of the earth, which is probably limited to five or six miles in depth, but that they diffuse themselves through the continuous unfractured zone down to depths where temperatures of fusion prevail, and that they there enter into combination with the lavas or with hot rock to form lavas. It is well known that aqueous vapor facilitates the fusion, or more accurately, the mutual solution of the minerals. This view is a part of one of the hypotheses concerning the origin of the lavas themselves.

The opposing view supposes that the gases were in the main original. Of this view there are two phases: (1) One supposes that the lavas are remnants of an original molten globe which absorbed gases from the primitive atmosphere and retained them till the time of their eruption. The possible absorption of steam and air into the supposed molten globe has been much neglected in current conceptions of early conditions. If the lavas of the supposed molten globe absorbed proportionately as much water-vapor as the volcanic lavas often contain, it would probably take forty or fifty times the present ocean and atmosphere to supply them. Any remnants of these original lavas might well be supposed to hold gases. Even rocks derived from them by deep-seated solidification might retain much gas. (2) The other phase of the view assumes that the gases were entrapped when the globe was built up of meteoroidal or planetesimal matter, as assumed in the accretion hypothesis.

Under either of the last two views the gases may be said to be primary, and genetically connected with the origin of the lavas themselves. Such gases would be a contribution to the atmosphere and the hydrosphere. This view does not exclude the idea that as the lava rises through the surface-rocks, other gases are formed by contact, and that they may be absorbed into the rising column. On the contrary, the view recognizes the possibility that a tongue of lava rising into the upper formations may encounter bodies of water, or masses of thoroughly water-soaked rock, from which great quantities of steam may be generated, and that this accessory steam may be a large factor in the initial explosion which often accompanies the development of a new volcano, or the new eruption of an old one after a long period of quiescence.

A decision on the vital question whether the volcanic gases are largely primary, or are essentially secondary, has not yet been reached; but it will doubtless be reached when a sufficient number of really representative analyses of volcanic gases have been made, and when the phenomena of the gases occluded in igneous rocks have been thoroughly investigated.

The peculiar proportions of the rock-gases, in which hydrogen and carbon dioxide so greatly preponderate, seem to imply that they are not derived from the atmosphere; at least if they were so derived, there must have been a selective absorption of a most remarkable kind, because hydrogen is present in the atmosphere in exceedingly small quantities, while carbon dioxide is a very minor constituent. At the same time, as already remarked, no free oxygen is usually found in these absorbed gases.

The question as to whether the larger part of the volcanic gases is original or is merely a special form of convective circulation, has an important bearing on the supply of the atmosphere, which is constantly being depleted by the oxidation of the rocks and by the formation of carbonates and carbonaceous deposits. This vital phase of the subject will receive further consideration. While recognizing the lack of decisive proof, it would seem that the preponderance of evidence lies in favor of the view that a notable portion, at least, of gaseous volcanic emanations is derived from the interior of the earth, and is really a contribution to the atmosphere and the hydrosphere. The hydrogen, on coming in contact with the atmosphere, ignites and adds itself to the hydrosphere. The carbon dioxide is in part decomposed by plants, and adds to the supply of atmospheric oxygen. The nitrogen, being comparatively inert, doubtless gradually accumulates in the air and has thus come to be its preponderant constituent.

THE CAUSE OF VULCANISM.

The extraordinary facts involved in volcanic phenomena cannot well be discussed fully until the origin of the earth is considered, and the great agencies, as well as the peculiar conditions, which the earth inherited from its birth, are duly weighed, for these were, with little doubt, the true causal antecedents of vulcanism. We shall return to the subject after a sketch of the early conditions of the earth, but the views that have been entertained may be reviewed here while the phenomena are fresh in mind.

The explanation of vulcanism involves two essential elements. These are (1) the origin of the lavas, which involves a consideration of the necessary temperatures, pressures, and other conditions, and (2) the forces by which the lavas are expelled.

Nearly all current explanations of vulcanism are founded upon conditions supposed to be derived from a molten globe, and fall under two general classes: (I) those which assume that the lavas are residual portions of the original molten mass, and (II) those which assign the lavas to the local melting of rock.

I. *On the Assumption that the Lavas are Original.*

In this case it is not necessary to assume any special accession of heat, but merely to account for extrusion. There are two phases of this view, (1) the one postulating a general molten interior, (2) the other limiting the molten matter to local reservoirs.

Hypothesis 1. Lava outflows from a molten interior.—In the early days of geology, when the earth was supposed to have a thin crust and a molten interior, it was very naturally assumed that volcanoes were but pipes leading down to the molten mass within. This view has been essentially abandoned. The independence of adjacent vents is in itself almost a fatal objection, when it is recalled that the height of recent volcanic craters ranges from nearly 20,000 feet above the sea, to 10,000 to 20,000 feet below. The view would involve the conception of lava-columns connected with a common reservoir varying possibly 30,000 to 40,000 feet in altitude, and certainly more than half that much, simultaneously. The lower outlets should as certainly be selected for the outflow of the great interior sea of fluid rock, as the lowest sag in the rim of a lake for its outflow, for no great differences in specific gravity are presumable under this hypothesis. An equally grave objection arises from tidal strain. If the earth were liquid within and merely crusted over by a shell of rock of moderate thickness, it would yield appreciably to tidal stresses, and this yielding would change the capacity of the interior so that with every distortion of the spheroid a portion of its fluid interior would be forced to the outside, and with every return to the more spheroidal form there would either be a re-flow to the interior or a shrinking of the crust. In any case a very demonstrative response to tidal influence would tell the story of interior fluidity. No such effects are observed. The tidal strains may perhaps have a slight effect in hastening a given eruption when the forces are approaching a delicate balance and an eruption is imminent, but the very triviality of this influence implies not only the absence of a general liquid interior, but also of extensive reservoirs.

Hypothesis 2. Lavas assigned to molten reservoirs.—A modification of the preceding view has been made to escape the difficulties involved in the hypothesis as stated above. It is supposed that while nearly all the subcrust solidified, numerous liquid spots were left scattered through it. This honeycombed substratum is supposed to connect the continuous outer crust with a central solid body, solidified because of pressure in spite of its high temperature. This hypothesis escapes only a portion of the objections. For instance, the lavas in Mauna Loa and Kilauea in Hawaii differ nearly 10,000 feet in height, and hence cannot well be supposed to connect with the same reservoir, but they are both on the same vast cone, which implies at least an equally large molten reservoir as its source. If there were two distinct reservoirs of the required magnitude, they must be singularly placed to supply vents so near and yet so independent. The difficulty grows greater when the whole Hawaiian chain is considered, for the points of eruption seem to have migrated from the northwesterly islands, where the volcanoes are old, to the southeastern end, where volcanic activity is now in progress.

It would be natural under this view to suppose that these residuary lakelets of liquid rock should be gradually exhausted as time goes on, and that vulcanism should be a declining phenomenon. It is not clear that this is the case. The great number of existing volcanoes in regions where great extrusions took place in earlier ages does not seem to be in harmony with the hypothesis.

II. *On the Assumption that the Lavas are Secondary.*

The serious difficulties that arise in interpreting volcanic lavas as remnants of an original molten mass, and the strong arguments of recent years for a very solid earth, have turned inquiry chiefly toward the second class of hypotheses, which refer the origin of lavas to the local melting of deep-seated rock. These differ widely among themselves. One group seeks for a cause of the melting in the penetration of surface air and water; another, in the relief of pressure; a third, in crushing and shearing; a fourth, in the depression of sediments into the heated interior zone; and a fifth, in the outward flow of deep-seated heat.

Hypothesis 3. Lavas assigned to the reaction of water and air penetrating to hot rocks.—As steam is one of the great factors in the explosions of volcanoes, and as water reduces the melting-point of rocks, it is a natural and simple view that water penetrating through the fissures and pores of the outer crust and coming into contact with the heated rocks below, is absorbed into them and renders them liquid, and that then, being rendered swollen and lighter by the process, they ascend and discharge quietly or explosively according to the special conditions of the case. Naturally the suggestion arises that the waters would be converted into steam long before they could reach rock hot enough to be melted, and that this steam would be forced back along its own track, as the line of least resistance, rather than force itself into the rock material and rise in the lava-column; but to this it is answered that an experiment of Daubree's has shown that water will penetrate the capillaries of sandstone against high steam pressure and add itself to the steam within. The fact is also cited that certain substances, when highly heated, absorb gases which they give out when they cool. The absorption of hydrogen by platinum, and of oxygen by molten silver, are illustrations. It is certain that the lavas do contain large quantities of absorbed gases, and that these are partly, and in most cases largely, given out in cooling, when the cooling takes place at the surface. The presumption is that the lavas would take the gases up again on remelting under similar conditions. If the lavas of actual volcanoes had the temperatures of aqueo-igneous fusion (700°–1000° Fahr.) only, it would strengthen this view; but as temperatures of lavas often exceed 2000° Fahr., and probably sometimes reach 2500° Fahr., and perhaps 3000° Fahr., it is not easy to account for such temperatures under this hypothesis, because they would only be reached at levels far below those at which aqueo-igneous fusion might be presumed to take place. Perhaps this could be met by invoking pressure which might prevent even aqueo-igneous fusion from taking place until these temperatures were reached, but pressure brings in a grave difficulty in another line, as we shall presently see.

There is a phase of the water-penetration hypothesis which seeks to account for an accession of heat. It is affirmed that the outer rocks are oxidized, while the inner ones were not originally, or at least not completely oxidized, and that air and water from the surface, reaching the unoxidized zone, enter into combination and generate the necessary heat. This view was pardonable before the development of modern thermo-chemistry, but is now quite untenable, as may be shown by working out the reactions thermally.

All views which assign the penetration of surface air or water as a cause meet with a grave, if not insuperable, difficulty in the condition of the lower part of the earth's crust (see p. 218). The fractured condition of the crust, which permits a ready penetration of water, is a very superficial phenomenon. Below the first few thousand feet the crevices and porosities of the rock are rapidly closed by the pressure of overlying rock, and all appreciable crevices and pores probably disappear at a depth of five or six miles. The effective function of fissures is, therefore, limited to the upper few miles of the crust, and even here to certain portions only. The great pressures in gas- and oil-wells show that in many quite superficial beds, even when arched, there are no fissures or pores capable of letting even gas escape effectively. The depths at which the temperatures of lavas are reached are usually estimated, from the downward increase of temperature, at 20 to 30 miles. This leaves from 14 to 24 miles of the compressed zone between the lowest assignable limit of the fissured zone, and the highest assignable zone for the origin of lavas. This thick zone of dense rock must be reckoned with in all hypotheses that involve the penetration of air and water from without, and, as well, the extrusion of lavas from within. In addition to the difficulties of the penetration of ground-water, the limitations of its heat, at penetrable depths, also bear adversely (see p. 219), on the descent of air and water.

Hypothesis 4. Lavas assigned to relief of pressure.—It seems to be demonstrated that pressure raises the melting-point of average rock, and hence at twenty or thirty miles' depth there may be rocks hot enough to melt at the surface, but still solid because of high pressure. If this pressure were in some way relieved they would become liquid. Pressure may be locally relieved somewhat (1) by denudation, (2) by certain phases of faulting, (3) by anticlinal arches, and (4) by continental deformation.

(1) In most cases of denudation, cooling below probably keeps pace with loss above. At any rate, many volcanoes rise from the bottoms of the oceans where no denudation takes place, and this phase of the hypothesis is not workable there.

(2) The theory of relief by faulting finds encouragement in the fact that many volcanoes occur on fault-lines. There is no evidence, however, that this is a universal or necessary relation. Computation as to the amount of lowering of the melting-point that might arise from the faulting associated with volcanoes indicates that it is necessary to suppose that the rocks were already very close to the melting-point when the faulting took place, to make the doctrine applicable. It is to be observed that in faulting the relief of pressure on one side of the fault-line is likely to be balanced by increased pressure on the other side, and that this difference in pressure may be lost by distribution at a depth of 20 or 30 miles, where, at the nearest, this delicate relation between solidity and liquidity, on which the theory is dependent, may perhaps be reached.

(3) Immediately under an anticlinal arch there may doubtless be some relief of pressure within the limits of strength of the arch, which is not great (p. 582). The pressure under the arch *as a whole* is greater than before it was bowed up by lateral thrust, and in depth this excess becomes distributed so as to obliterate the local relief under the center of the arch, and so *adds* the effects of folding to the average pressure of the crust. Besides, as a matter of fact, volcanoes do not appear to be especially associated with mountain folds where arching reaches its best expression.

(4) The same general considerations bear on the assignment of liquefaction to relief of pressure in connection with the more general deformation of the earth's body. Besides, while relief of pressure might account for liquefaction, it leaves the extrusion without an obvious cause; indeed, it would seem to furnish a condition opposed to extrusion, and if pressure were subsequently added to force the liquid out, it would tend to restore the solid condition.

Hypothesis 5. Lavas assigned to melting by crushing.—Mallet[1] and others have attributed melting to the crushing of rock. Crushing, in the ordinary sense of the term, can only take place in the zone of fracture, and that is apparently too shallow to meet the requirements of the case. Below this zone, the pressure on all sides is too great to permit any separation of fragments, and a solid mass can only change its form by what is called "solid flowage." The rock under these conditions may be compressed, and this compression must give rise to heat, but at the same time the melting-point is raised, according to all experiments. It seems improbable that melting can be produced in this way. If great pressure could be brought to bear upon a tract of rock so as to heat it by compression, and if then the pressure were relaxed before the heat generated could be distributed by conduction, and if re-expansion did not follow, possibly melting could be effected, but this makes the process complicated and apparently inapplicable. It is scarcely possible that such a sequence of events can have affected all the tracts that are now volcanic, much less all those that have been such throughout geologic history. As noted in the preceding case, relaxation would seem to be unfavorable to expulsion. Besides, volcanoes do not seem to be confined to tracts that show signs of great crumpling and crushing, as the Alps, the Appalachians, and the closely folded ranges generally.

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[1] Phil. Trans., 1873.

Extrusions seem rather more common with faulted ranges where crushing is less notable and where surface tension replaces compression.

Hypothesis 6. Lavas assigned to melting by depression.—It is observed that in certain regions great thicknesses of sediments have accumulated by the slow settling of the crust below, and as these sediments obstruct the outward flow of heat while the lower beds settle nearer to the interior source of heat, it is conceived that they become heated below and, being saturated with water, take on aqueo-igneous fusion and rise as lavas, well supplied with internal gas and steam from the water and volatile constituents that were entrapped and carried down with them. The question obviously arises whether such depression is sufficient to give the temperatures the lavas show, and whether volcanic action is confined to such areas of depression and deep sedimentation. At the highest credible estimates—which are none the less to be taken with reserve—the post-Archean sediments rarely reach five or six miles in thickness at any given point, and probably never exceed ten or twelve, while twenty or thirty miles is the computed depth required for the acquisition of the temperatures the lavas actually possess. If the Archean terranes be included among the sedimentaries, the thickness may be adequate, but what then of the Archean vulcanism, which much surpasses that of later times, and the other early extrusions before the sediments were thick; and what of the moon, where there are probably no sediments at all?

Besides, it is not at all clear that the distribution of vulcanism is specially related to that of thick sediments, as it should be if this hypothesis were the true one. There are many volcanoes in the heart of the great oceans where sedimentation is now inappreciable, and probably has been in all past periods.

Hypothesis 7. Vulcanism assigned to the outflow of deep-seated heat.—If the earth grew up by slow accessions of meteoroidal or planetesimal matter, in a manner to be more fully set forth in the discussion of the origin of the earth, and if its interior heat be due chiefly to compression by its own gravity, the internal temperature would be originally distributed according to the degree of compression, and this would depend on the intensity of the internal pressure. This can be approximately computed, and is shown in the diagram on page 563, where this subject has been treated. On not improbable assumptions regarding the thermometric conductivity, the flow of heat from the deep interior to the middle zone would be greater than the loss of this zone to the superficial zone. This middle zone should, under this view, experience a rising temperature. By hypothesis, this zone is composed of various kinds of matter mixed as they happened to fall in. Hence as the temperature rises, the fusion-points of some of these constituents will be reached before those of others. More strictly, the temperatures at which some of these constituents will mutually dissolve one another will be reached, while other constituents remain undissolved, and thus a partial and distributed liquefaction will arise. The gases and volatile constituents in the mixed material would naturally enter largely into the liquefied portion. It is assumed that with a continued rise of temperature, the partial liquefactions would increase until the liquefied parts found means of uniting, and the lighter portions, embracing the gaseous contingent, were able to work their way toward the surface. As they rose by fusing or fluxing their way, the pressure upon them became less and less, and hence the temperature necessary for liquefaction gradually fell, leaving them a constantly renewed margin of temperature available for melting their way through the upper horizons. Thus it is conceived that these fusible and fluxing selections from the middle zone might thread their ways up to the zone of fracture and thence, taking advantage of fissures and fractures, reach the surface. It is conceived that such liquefaction and extrusion would carry out from the middle zone the excess of temperature received from the deeper interior, and thus regulate its temperature and forestall general liquefaction, the zone as a whole remaining always solid. The independence of volcanoes is assigned to the independence of the liquid threads that worked their way to the surface. Nothing like a reservoir or molten lake enters into the conception. The prolonged action of volcanoes is attributed to the slow feeding of the liquid threads from the locally fused middle zone. The frequent pauses in action are assigned to temporary deficiencies of supply; the renewals to the gathering of new supplies after a sufficient period of accumulation. The distribution of volcanoes in essentially all latitudes and longitudes is assigned to the general nature of the cause. The special surface distributions are assumed to be influenced, though not altogether controlled, by the favorable or unfavorable conditions for escape presented by the crustal seg- ts of the earth. The persistence of volcanic action in time is attributed to the magnitude of the interior source, to its deep-seated location, and to the slowness of conduction of heat in the earth's interior. The force of expulsion is found in the stress-differences in the interior, particularly the periodic tidal and other astronomic stresses (see p. 580), and in the slow pressure brought to bear on the slender threads of liquid by the creep of the adjacent rock. The violent expulsions are due to the included gases, of which steam is chief. Little efficiency is assigned to surface-waters, and that little is regarded as wholly secondary and incidental. The true volcanic gases are regarded as coming from the deep interior and as being true accessions to the atmosphere and hydrosphere. The standing of the lavas in volcanic ducts for hundreds and even thousands of years with only small outflows, as in some of the best-known volcanoes, is regarded as an exhibition of an approximate equilibrium between the hydrostatic pressure of the deep-penetrating column of lava, and the flowage-tendency of the rock-walls, the outflow being, of course, also conditioned on the slow rate of supply below, and the periodic stress-differences of the interior.

For the present these hypotheses must be left to work out their own destiny, serving in the mean time as stimulants of research. All but the last have been for some time under the consideration of geologists, and are set forth in the literature of the subject (p. 636).

A few special phases of the problem need further discussion, though they have been incidentally touched upon.

Modes of Reaching the Surface.

All of the views that locate the origin of the lavas deep in the earth must face the difficulties of the passage through the dense portion of the sphere below the fracture zone. Near the surface, the lavas usually take advantage of fissures or bedding-planes already existing or made by themselves. There is little evidence that they bore their way by melting, though they round out their ducts into pipes as they use them, much as streamlets on glaciers falling into crevices round out moulins. But this use of fissures and bedding-planes for passage is probably merely a matter of least resistance where the lavas are relatively cool, and their capacity for melting is low or perhaps even gone. Daly has recently urged that lavas work out reservoirs and enlarge passageways for themselves by detaching masses of rock from the roofs and sides of the spaces already occupied by them, these masses either melting and mingling with the lava, or else sinking to lower positions in the column. This process he designates *stoping.*[1]

In the denser and warmer zone below, the alternatives seem to be (1) melting or fluxing, or (2) mechanical penetration without fracture. As rocks "flow" in this zone by differential pressure without rupture, an included liquid mass may be forced to flow through the zone by sufficient differential pressure. If local differential pressures at the surface be neglected as probably incompetent, there only remain the stress-differences of the interior and the differences of hydrostatic pressure between the lava-column and the surrounding solid columns. The latter would not be great until a column of liquid of much depth was formed, and the former would probably not be concentrated on the liquid in such a way as to force it bodily through the solid rock. Probably fusing or fluxing its way with the aid of stress-differences is the chief resource of the lava in the initial stages. In this it may be supposed to be assisted by its gases, by its selective fusible and fluxing nature, by its very high temperature if it comes from very great depths, as held in the seventh hypothesis, and by the stress-differences which prevail in the deep interior, as shown in the last chapter. In ascending from lower to higher horizons, the lava would be constantly invading regions of lower melting-point because of less pressure. It would thus always have an excess of heat above the local melting temperature until it invaded the external, cool zone, where the regional temperature is below the melting-point of surface pressure. From that point on it must constantly lose portions of its excess of temperature by contact with cooler rocks, and probably in the process of fluxing its way in the compact zone. If this excess is insufficient to enable it to reach the zone of fracture, the ascending column is arrested and becomes merely a plutonic pipe or mass. If it suffices to reach the zone of fracture, advantage may be taken thereafter of fissures and of rupturing, and the problem of further ascent probably becomes chiefly one of hydrostatic pressure, in which the ascent of the lava-column is favored by its high temperature and its included gases. The hydrostatic contest is here between the lava-column, *measured to its extreme base*, and the adjacent rock-columns, *measured to the same extreme depth*. The result is, therefore, not necessarily dependent on the flowage of the outer rocks, but may be essentially or wholly dependent on the deep-seated flowage of the rock of the lower horizons. The ascending column may reach hydrostatic equilibrium before it reaches the surface, and may then form underground intrusions of various sorts without superficial eruption, or it may only find equilibrium by coming to the surface and pouring out a portion of its substance and discharging its gases.

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[1] Mechanics of Igneous Intrusion, Am. Jour. Sci., Apr., p. 269, and Aug., p. 107, 1903.

Additional Considerations Relative to the Gases.

The question whether the volcanic gases are a contribution to the atmosphere and hydrosphere is so important in its bearings on the whole history of the atmosphere as to merit additional consideration here. As already noted, if the volcanic gases arise from water and absorbed air that have previously passed down through the strata, there is no real contribution to the hydrosphere and atmosphere, but merely circulation. If the gases are chiefly derived from the deep interior, they are an important accession to the atmosphere and hydrosphere.

Most views are more or less intermediate, assigning a part of the gases to the interior and a part to the exterior. No one will question that some part at least of the steam is due to the contact between the ground-waters and the hot lava, and probably no one will question that some gas comes from the interior if the lavas originate there. The vital question is, whence comes the major portion? Are the constant ebullitions of some volcanoes and the terrific explosions of others due mainly to surface-waters, or to interior gases?

It seems to be certain that in most cases the gases are diffused through the substance of the lava, and are not simply in contact with the walls of the column or with its summit. Without doubt steam is generated around the lava-column by external contact, and perhaps some explosions are due to the entrance of the rising lava upon a crevice or cavern filled with water, or to the invasion of a lake gathered in an old crater; but it still remains a question whether the importance of such explosions has not been exaggerated. Such action does not seem competent to produce *inflated* lavas, but merely shattered ones. Water thus "suddenly flashed into steam" could scarcely diffuse itself intimately through the lava, for the process of diffusion is exceedingly slow. But inflated lavas, pumice, scoriæ, and cinders are the typical products of explosive vulcanism. Not only in the ordinary Vesuvian type, but in the extraordinary Krakatoan type, inflated lavas are the dominant product. Prodigious quantities of this covered the sea about Krakatoa after its tremendous explosion in 1883. Judd estimates that the volume of included steam involved in the inflation of the pumice examined by him, was from three to five times that of the rock, and that the amount involved in exploding the lava into the fine dust that floated in the upper atmosphere for months, was presumably much greater.

If the sudden flashing into steam of bodies of water in external contact with hot lava be rejected as only an incidental source of explosion, it remains to be considered whether waters permeating the rock and becoming converted into steam may not be absorbed into the rising lava, become diffused through it, and ascending with it, explode at the surface. So far as access through fissures and cavities large enough to be entered by lava are concerned, it may safely be concluded that since the hydrostatic pressure of the lava must be greater than that of the water in the fissures, or else it could not rise, the lava will enter them, forcing back the water or the steam generated from it, and, having penetrated as far as accessible, will solidify as a dike, and plug up the avenue of contact between the ground-water and the portion of the lava still remaining molten. The numerous dikes that attend volcanic necks testify to the prevalence of this action. The capillary pores of the wall-rock, which cannot be thus bodily occupied by the lava, must doubtless become filled with steam, and this, following the principles of Daubree's experiment, will force itself into contact with the lava and be absorbed by it, but whether this will be in sufficient quantity, and will become sufficiently diffused through the body of the lava-column to produce the observed effects, is an open question. The increasing testimony of deep mining is that relatively little water flows through the deeper horizons. It is urged that the water remaining in solidified lavas is very unequal in distribution, as though due to unequal access and partial diffusion. The argument seems strong, but to make it thoroughly good, it must be shown that this inequality is not due to *irregularity of discharge* of the gases during and after eruption, rather than to irregularity of original *accession*. There is, perhaps, as much ground for assigning differences in the degree of parting with the included gases, as in acquiring them. Doubtless those lavas that boiled and seethed for a long period in the caldron were more fully deprived of eir gases than those that were more promptly disgorged and cooled th less convection and surface exposure.

Thermal considerations.—Probably the most important consideration relates to the heat effects. If underground-waters enter the lava-column and come forth as steam, great quantities of heat are consumed in the process. Has the lava a sufficient excess of heat to stand this? Can ebullition be maintained for the observed periods if the steam comes from ground-waters?

Many lavas probably do not carry a very large excess of heat above that necessary for liquefaction, for not a few of them contain crystals already forming, which shows that they are within the range of the temperatures of solidification of their constituents. The same conclusion is indicated by the limited fusing effects shown by the walls of dikes and sills. On the other hand, as already remarked, dikes and sills often show the effects of a rather rapid cooling from the walls. The method of flow often implies the same condition for the acidic lavas, since they usually behave as stiff, pasty masses of limited liquidity. On the other hand, the basic lavas, whose fusion-point is much lower, often flow freely and reach great distances before solidifying. The facts taken altogether imply that the average temperature of the lavas is not much above the fusing-point of the acidic lavas, while it may probably be very considerably above the fusing-point of basalt. For a rough estimate, with no pretensions to accuracy, it may be assumed that in an average case there are 500° Fahr. excess, but probably not 1000° Fahr. A computation based on even so rough an estimate as this may, by showing the order of magnitude of the thermal considerations, indicate their radical bearing. The average temperature of the ground-water of the upper two or three miles of the crust—the only portion through which water probably penetrates with sufficient freedom to be effective in this case—is probably less than 200° Fahr. The specific heat of rock appears to average somewhat less than 0.2. The temperature of the lava may be taken at 3000° Fahr. as a sufficiently high average. From these data it follows that if an amount of ground-water equal to five percent. of the volume of the lava entered the lava and was brought up to its temperature and then discharged, the temperature of the whole mass would be lowered 550° Fahr. It is therefore evident that only a small percentage of surface-water can pass through the lava consistently with its continued fluidity.

M. Fouqué estimated that the discharge of steam from a merely *parasitic cone* of Etna during 100 days was equal to 2,100,000 cubic meters of water. If this were ground-water, and the lava from which it issued had an excess of 500° Fahr. above the fusion-point, the formation of this steam would congeal a column 400 feet in diameter and 3000 feet deep in the time given. If this case is typical, and if Fouqué's estimate is not greatly exaggerated or very exceptional, the view that any large portion of the steam from volcanoes comes from surface-waters seems to be incompatible with the persistence of ebullition and explosion which many of them exhibit. Stromboli has been in constant eruption as far back as the history of the region runs. It is now exploding every three to ten minutes, and yet the mass of lava seems to be small and its outflow inconsiderable. Is it possible that a current of steam, given out with this activity for so long a period, was derived from adjacent ground-waters, and has not yet solidified the lava?

The problem takes on a very different aspect if the steam, or at least some large part of it, rises from great depths and brings thence an excess of heat. It then becomes an agency for the maintenance of the liquidity of the lava, for giving it convective motion, and for promoting explosive action, so long as it continues to rise.

For these and other reasons the balance of present evidence seems to us to favor the view that most of the steam and other gases come with the lava from its original source deep in the earth.

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CHAPTER XI.

THE GEOLOGIC FUNCTIONS OF LIFE.

I. THE DISTINCTIVE FEATURES OF ORGANIC PROCESSES.

THERE is no reason to suppose that life processes, as we know them, were in operation in the earliest stages of the earth's history. They were introduced and developed gradually during its progress. With life there came into the processes of the earth's development three distinctive factors:

A. Certain chemical actions giving rise to compounds that are not known to occur independently of life.

B. Certain modes of aggregation of material, and certain kinds of bodily movements, not known except in association with life.

C. The mental element, under the direction of which certain new processes were inaugurated, and certain previous processes were modified and controlled.

A. THE CHEMICAL WORK OF LIFE.

The peculiar chemical phenomena connected with life chiefly concern the carbon compounds. In the inorganic world the carbon compounds are few and simple. In the organic world they become extremely numerous and complicated. These compounds are very unstable, for the greater part, and their partial decomposition gives rise to many additional compounds. Some of the true organic compounds and some of their decomposition products have the power of combining with inorganic substances, and so produce an additional series of semi-organic combinations. The total number of the compounds thus directly and indirectly connected with life greatly exceeds that of all inorganic compounds. Their mass, however, is very greatly inferior.

Life material chiefly atmospheric.—In the building up of the organic compounds, a necessary step is the decomposition of certain inorganic ompounds. The chief of these is the carbon dioxide of the atmosphere d hydrosphere, the decomposition of which furnishes the carbon eded for the organic compounds. On this account carbon dioxide may be regarded as in some sense the basal material or the fundamental food of the organic kingdom, and hence it plays a radical rôle in the life-history of the earth.

Water, and the constituents of water, oxygen and hydrogen, play a larger part quantitatively, but a less distinctive part.

Nitrogen is also an essential element, and usually stands next to carbon, oxygen, and hydrogen in quantity.

These, it will be noted, are all atmospheric constituents, and the material of life is, therefore, dominantly atmospheric. This is even true of aquatic life, for it lives largely on the atmospheric constituents dissolved in the water. The function of life, considered from the material point of view, is not only fundamentally concerned with the atmosphere, and intimately dependent on its conditions, but its most important material effects appear to lie in its modification of the constitution of the atmosphere.

The non-atmospheric factors.—The atmospheric constituents are not, however, the only elements intimately connected with the life function. Compounds of sulphur, phosphorus, potassium, sodium, chlorine, iron, calcium, magnesium, silicon, and other elements are more or less essential to the life of many organisms, or are employed by them for their skeletons, coverings, etc. Incidentally, nearly all the common elements become intimately related to living organisms either in the relations of active elements in their physiological functions, or of passive elements in their structure or in their auxiliary parts.

Three Classes of Effects.

Out of life processes grow three rather distinct classes of results: (1) changes in the amounts and proportions of the constituents of the atmosphere and, to some slight extent, of the hydrosphere and lithosphere; (2) aid or hindrance to inorganic processes, such as disintegration, erosion, and deposition; and (3) distinctive products, either (*a*) of organic matter that would not have come into the existing combination but for life, such as peat, lignite, amber, etc., or (*b*) of special forms of inorganic matter that would not have arisen but for life, such as coral deposits, shell-marl, diatom ooze, etc.

(1) *Changes in the composition of the atmosphere.*

The succession of modifications which the atmosphere has undergone from time to time through the action of life will be discussed as the earth's history is followed in the second volume. It may suffice here to note briefly the chief ways in which the atmosphere has probably been modified by the agency of life, not only as regards its quantity, but also as regards the proportions of its constituents.

The consumption and restoration of carbon dioxide.—As the fundamental food of the organic world, carbon dioxide has suffered enormous consumption in the course of the geological ages, and is now reduced to the very small proportion of .0004 or .0003 of the whole. At the outset it was probably one of the most abundant constituents; possibly even the chief one. It has been partially restored, concurrently with its consumption, by animal respiration, by certain classes of plant action, and by combustion and other forms of inorganic combination. This restorative action has been incomplete at all known stages of the earth's history, and hence there has been constant loss of carbon dioxide. The inorganic processes which have also profoundly affected both the consumption and restoration of carbon dioxide are here neglected and discussed elsewhere.

The freeing and consumption of oxygen.—The oxygen of the atmosphere is actively consumed by animals and by plants, but on the other hand, it is set free abundantly by green plants, and hence its amount has probably fluctuated from time to time according to the state of balance between the organic processes of its production, and those of its consumption. The consumption of oxygen by organic processes is, however, little more than a reversal of the previous process by which it was set free; for instance, green plants in forming their food set free the oxygen of the carbon dioxide used for the purpose. When the organic substance so formed is ultimately consumed through plant or animal action or by inorganic means, an equivalent amount of oxygen reunites with the carbon to again form carbon dioxide. And so if the whole of the organic matter is returned to the inorganic state, no more oxygen is consumed than had been before set free in the process of forming the organic matter. But, as a matter of fact, a large amount of organic matter has not gone back completely to the inorganic state, and this residue constitutes a factor of no small importance in the geological record.

The organic residue.—There is a certain portion of vegetation that is not consumed by animals or by other plants, and that escapes combustion and all kinds of ordinary decay, and this constitutes a part of the organic residue. Animals never completely oxidize all the organic matter they take into their systems; their bodies never entirely consume themselves. A like statement may be made respecting those plants that feed on organic matter. That which animals and plants leave unoxidized is indeed more or less preyed upon by other animals and plants, and relatively little escapes final reoxidation, but there is a remnant, and this constitutes another part of the organic residue. The more conspicuous forms of the organic residue are found in the mucks, peats, lignites, coals, organic oils, and gases, but in addition there is not a little disseminated organic matter in nearly all the sedimentary rocks; in the aggregate, this probably amounts to more than the distinct organic deposits.

The meaning of the organic residue.—All the unoxidized, or incompletely oxidized, carbon in the organic residue implies that oxygen has previously been separated from this residual carbon by plants and given to the atmosphere, and hence has been a source of atmospheric enrichment in oxygen. The amount thus contributed is equal to that which is required to restore the residual carbon to its original state of oxidation. So, in a similar way, the unoxidized hydrogen in the organic hydrocarbons and like compounds implies that oxygen has been separated from the hydrogen of water and given to the atmosphere, and hence this also is a source of atmospheric enrichment in oxygen. It seems safe, therefore, to conclude that the action of life, taken as a whole, has increased the free oxygen of the atmosphere.

While not here under consideration, it is not to be forgotten that inorganic processes involving the same atmospheric constituents have been in operation concurrently with the organic processes, and that they have also affected the amounts and proportions of the atmospheric constituents. Rocks have been oxidized in greater or less measure at the expense of the atmospheric oxygen, and hence when the total atmospheric problem is considered, there arises the question whether the amount of oxygen in the atmosphere has been increased or diminished during geological history, when the balance is struck between the inorganic and the organic actions. The probabilities seem to us to strongly favor the view that organic action has preponderated, and that the oxygen has been increased beyond its primitive amount, but that it has fluctuated during known geological history. The reasons for this view will appear in the historical chapters.

The disintegration of the crystalline rocks and the solution of limestone have consumed much carbon dioxide, and this is to be added to the loss through organic action. On the other hand, there are inorganic processes that supply carbon dioxide, and hence when the larger problem of the atmosphere is raised, the factors become so complicated that their consideration is best deferred to the historical chapters. This passing reference may stand us in good part lest we forget, for the moment, the inorganic factors in the atmospheric problem.

The more inert factor.—Nitrogen in the free state is relatively inert chemically, and it does not appear that it can be used directly by the higher plants and animals in appreciable amounts. Certain bacteria, and perhaps certain algæ [1] and other low forms of plants, have the power of using free nitrogen, and this is a principal way in which it is put within the reach of higher plants. Nitrogen is also combined in small quantity in the atmosphere by electric action, and thus made available for plants. On account of the inertness of nitrogen and of the relatively limited amount required for organic purposes, the nitrogen of the atmosphere has been less consumed than the carbon dioxide. Besides this, the nitrogen compounds are very decomposable, and are very generally and completely returned to their original state. Deposits of nitrates or other nitrogenous compounds are relatively rare.

It is obvious that if there is any considerable source of supply concurrent with this slight loss, the amount of nitrogen in the atmosphere must have been increasing. We have seen that volcanoes give forth considerable quantities of nitrogen, and that this *may* be a real addition to the atmosphere, and not merely a return of the atmospheric nitrogen that had been carried down previously by underground-water. It has also been noted that crystalline rocks contain occluded nitrogen, which is doubtless freed by their disintegration. It is, therefore, not improbable that the nitrogen of the atmosphere has been increasing, both actually and relatively.

Probable fluctuations of atmospheric composition.—With this general sketch of the interplay of the atmospheric elements under organic influence, we are prepared for the further conception that if one or another of these actions was relatively more vigorous than usual for a period, it would bring about a variation in the proportions of the atmospheric constituents. If, for example, vegetation flourished luxuriantly for a long period, but was measurably protected from the organisms that preyed upon it and from inorganic decomposition, as by falling into water or by prompt burial under sediment, the atmosphere might be growing richer in oxygen. If, on the other hand, vegetation were being relatively reduced, as *perhaps* it is being reduced now by man, and if previous organic products were being reoxidized at an unusual rate, as they are now in the burning of timber, coal, natural oil and gas, the carbon dioxide of the atmosphere might be relatively increasing, while the oxygen might be relatively diminishing. The possible fluctuations of the atmosphere as the result of organic action are, therefore, matters of vital importance, and invite attention in the historical study of the earth and in the outlook into its future.

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[1] Frank. Lehrbuch der Botanik, I, p. 576, 1892.

The climatic effects of organic action.—Interest does not, however, rest at this point. The researches of physicists have made it probable, if they have not altogether demonstrated, that the composition of the atmosphere has much to do with the climatic conditions at the surface of the earth. The atmosphere blankets the earth and equalizes its temperature. Acting as a screen, it subdues in some measure the intensity of the sun's rays by day, while it retards the radiation of the earth's acquired heat at night. This is in some measure the function of all the constituents of the atmosphere, but by no means of all equally. The oxygen and nitrogen are relatively diathermous, letting the sun's rays pass in freely, and the earth's rays pass out freely; but carbon dioxide and the vapor of water are much less diathermous, particularly to rays of low intensity, such as are thrown out by non-luminous bodies like the earth. It follows that while the solar rays come in rather freely and heat the surface of the earth, the dark rays which the earth radiates back are measurably arrested by the carbon dioxide and vapor of water, and serve to keep the air warm. The influence of the vapor of water is vividly shown in the different degrees to which cooling takes place at night in a dry and in a moist atmosphere, respectively, where other conditions are the same. Ice is said to form at night in desert regions where the air is extremely dry, even within the tropics, while in humid regions of the same latitude and altitude oppressively hot nights are common. The influence of the carbon dioxide is not thus familiarly demonstrated, since its amount varies but slightly in different localities, but physical experiment indicates that it has a similar function.

If the amount of carbon dioxide in the atmosphere varies from age to age, the climate of the earth must apparently vary accordingly, and on this is built one of the hypotheses of climatic variation subsequently to be considered. We shall find that there have been great changes in the climate of the earth during its history. There is good evidence of former glaciation, not only in the northern United States and in England, Germany, and central Russia, but in India, Australia, and South Africa. At other times, figs and magnolias grew in Greenland and Spitzbergen, and corals flourished in the Arctic seas. There is good evidence of arid periods where humidity now prevails, and of humid periods where aridity now prevails. It is not assumed that the influence of organic action on the atmosphere has been the sole, or perhaps even the main, cause of these great climatic changes, but it is believed that it has been an important contributing factor. It is even possible that the climate of the future is much dependent on the agency of man, as implied above, however little ground there may be to suppose that he will, with altruistic purpose, control his action with a view to its bearing on the generations that may live tens of thousands of years hence.

(2) *Aid and hindrance to inorganic action.*

The promotion of disintegration.—While the influence of organic action on the lithosphere is quite superficial, and far less radical than that on the atmosphere, it is still important. Plants promote both disintegration and disaggregation under certain conditions, and hinder them under others, as already set forth. Chemical action of a decomposing and solvent nature takes place in connection with the roots of plants, while their growth sometimes rends rocks into whose crevices they have insinuated themselves. The acids and other products of organic growth and of organic decomposition attack some of the constituents of the rocks and contribute to their solution and disintegration. On the other hand, organic matter entrapped in the sediments, and so introduced into the strata at various depths, often acts as a reducing agency, causing the deposit of substances carried in solution in the underground-waters. Ores are sometimes thus formed, as explained in the discussion of ore-deposits (p. 476). Organic action on the whole promotes solution and disintegration at the surface, and prepares the way for deposition below.

Protection against erosion.—Another important function of vegetation is the protection of the land surface against erosion, as already noted in the discussion of erosion. A mantle of grass, especially if it forms a turf, or a carpet of leaves protected by bush and forest, greatly retards surface wash. It does this not only because it directly covers the soil, but because it holds back the run-off and tends to prevent those violent floods which give to erosion its greatest intensity. There is a marked difference between the erosive work which a given amount of water will do if, in the one case, it runs off gradually, and in the other, precipitately. By way of offset, it is to be noted that the disintegrating action of vegetation prepares the rock material for easy erosion, and to this extent helps in its removal by the drainage; but on the average this is greatly overbalanced by the protection afforded by the vegetal covering, though this is not true in every instance.

The influence of land vegetation on the character of the sediments.— The presence or absence of a vegetal covering influences the kind of deposit which is derived from the land, particularly if the surface be occupied by crystalline rocks. If the surface be well clothed with vegetation, the crystals of the complex silicates, such as the feldspars, micas, and ferromagnesian minerals, are usually disintegrated into clayey products before they are removed, so that, when borne away and deposited, the result is common shale. Concurrently, the relatively undecomposable quartz-grains are rounded into sand, and deposited as common quartzose sandstone, while the calcareous material is borne away in solution and deposited as limestone. But if the surface be bare of vegetation, the crystalline rocks are usually *disaggregated* before they are decomposed, for destructive action works best at the junctions of crystals, and along cleavage lines, and hence the crystals are usually separated from one another before they are fully decomposed. In the absence of a covering to hold them in place until they are decomposed, they are apt to be washed away, and the resulting deposit consists in considerable part of grains of feldspar, mica, hornblende, and other minerals, which do not usually occur in well-decomposed sediments. The deposits are, therefore, of the nature of arkose, if the original rocks are granitic, or of the nature of wacke, as the term is used in this book, if they are of the basic type. On this is based the inference that a vegetal covering of the land extended as far back in the history of the earth as clay shales, quartzose sandstones, and limestones form the prevailing sediments.

(3) *Distinctive deposits.*

Organic rocks.—In the chapter on the origin and descent of rocks, a group of rocks formed directly from organic matter is recognized and described. The chief of these are peat, lignite, bituminous coal, anthracite, and graphite. It is the belief of many geologists that natural gases, oils, and asphalts are also mainly derived from animal and vegetal remains. An alternative view, advocated by Mendelejeff and Moissan, assigns the oils, gases, etc., in part at least, to deep-seated carbides to which water has gained access and developed hydrocarbons, after the analogy of acetylene.[1] Whatever may be the truth relative to inorganic action, it is clear from geological conditions that some of the natural gases and oils are organic products. Besides the more common organic deposits, there is a long list of minor products, among which are amber, copalite, paraffine, ozocerite, camphene, etc. Guano and coprolites represent the excrementitious class.

Inorganic rocks due to life.—Besides these deposits of organic matter, or of its decomposition products, there is a large class formed from the inorganic matter that served auxiliary functions in the economy of life, such as shells, skeletons, etc. For the greater part these are composed of calcium carbonate, and give rise to limestones, marls, chalk, etc. Not a few, however, are silicious, and give rise to flints, cherts, and silicious earths. Some are formed of calcium phosphate, and a few of other inorganic material. The deposits formed in these ways have been defined in the chapter on rocks.

Fossils.

The term *fossil* is used so comprehensively as to include not only the remains of plants and animals themselves, but their tracks, impressions, casts, replacements, and all other distinct traces. It also embraces nests, borings, implements, and other distinctive products. These enter into the formation of the two classes of rocks just considered, but they have an independent function. They constitute the specific record of life, and their study not only reveals much of the past history of plants and animals, but furnishes one of the most important means by which the ages of formations are determined. In the early development of the science it was found that the uppermost and hence the latest beds of rock contain fossil forms either identical with those now living, or closely similar to them; that beds below these bear life relics that depart somewhat more from the living forms, and are somewhat less highly developed; that beds still lower bear fossils that depart still more from the living types, and are more primitive in general, and so on down as far as fossils are found.

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[1] Science, Vol. VI, p. 838, 1897. Zeitschrift für Anorganische Chemie, 1897.

The general order of life succession determined by stratigraphy.—Thus it appeared from the evidence of the strata that there was a general order of life succession. It was also found that this was, *in its main features*, the same for all the continents. By continued and close studies, the particulars of the succession were worked out more and more fully, and the work is still being pushed forward to greater and greater degrees of refinement. At the same time, it was found that there were different faunas and floras in different parts of the world in past times, much as there are now; that there were shiftings and migrations as now; that given species were increasing in some regions and dying out in others, and that innumerable variations and complications entered into the evolution and distribution of the life forms. But under and through all these there run a sufficient number of common features to show beyond reasonable question the order of succession of life.

Throughout all this study, *the chief guide was the actual order* in which the fossils were found in the succession of strata, because there is no evidence so conclusive of the order of events as the superposition of the sedimentary beds when they are normal and undisturbed. By the study of the fossils in the successive beds, it was found that there was a more or less progressive evolution of plants and animals brought about by modifications of their forms, and that these modifications assisted in determining the order of succession when the evidence of the strata was defective; and so the biological and stratigraphical factors reacted helpfully on each other.

Fossils as means of correlation.—While stratigraphy was thus, in the earliest stages, the main reliance in determining the order of events, and biology was the chief gainer, in the end stratigraphy received ample compensation, if indeed it did not become the greater beneficiary; for at no known and accessible place is there a complete succession of sedimentary beds. There are great series here and there, but their connections with one another are more or less concealed by surface formations or water-bodies. So also at many places the stratified series has been broken up by deformation, or cut away by erosion. Hence there was need for some reliable means of matching the beds of separated series, and of making up a complete ideal series. This means is found in the fossils they contain. While the variations of the faunas and floras in different regions, and their migrations, introduce some minor difficulties, the relations of the fossiliferous beds of one region to those of another can be determined with great satisfaction, and often with great precision. This is particularly so when abundant floating or free-swimming species lived in the seas and were freely fossilized, for they were deposited on the coasts of all the continents at practically the same time, and no uncertainties from migration or local differences in rate of evolution intervened to throw doubt upon the correlation. Without the aid of fossils, the correlation of the deposits on the separate continents would be attended with grave obstacles and much uncertainty, if not with quite prohibitive difficulties.

B. SPECIAL MODES OF AGGREGATION AND OF MOVEMENT.

Inorganic solid matter is chiefly crystalloidal; organic matter is chiefly colloidal; but there are colloidal states of inorganic matter and there are crystalloids among the organic products. In the inorganic world, solids very generally tend to organize in the form of crystals; in the organic world, they as generally tend to organize in the form of cells. Neither tendency is complete or exclusive, but each is dominant in its own sphere.

Still more distinctive than the formation of cells is the growth of complex organized bodies, the differentiated members of which perform special functions for one another, and are mutually dependent on one another. This is a profound departure from the habitual modes of the inorganic world.

Still more so is the power of voluntary motion in more or less disregard of outside physical influences. Through this power, distribution may take place contrary to current and wind, and to gravitation itself. From the view-point of past geologic transportation, this is perhaps more singular than important, for no great mass of matter has been transported contrary to the influences of gravity, wind, and current, by the exercise of this peculiar power of animals, but it is not without geologic importance in the migrations and in the redistributions of organic influences that arise from migrations. When the influence of man is included, the geologic effects require consideration, but here the third distinctive factor, the mental element, comes into effective play, and we pass to its consideration.

C. The Mental Element.

Current opinion does not recognize a mental element as residing in the plant world, and it is divided as to the degree of its development in the lower animal kingdom, but its influential presence in the higher animal orders and in man is beyond legitimate question. Two phases are to be recognized: (1) the material work done under the stimulus and direction of mental impulses, as, for example, excavations, transportations, changes of drainage, removal of forests, cultivation of soil, etc., and (2) the intellectual work of the faculties themselves irrespective of material changes. In one view, geology is a purely material science concerned solely with the formation of the earth and with the physical development and relations of its inhabitants. In another, geology is a comprehensive historical science concerned with every phase of the world's history, and certainly not least with the higher forms of life development, with their psychological, sociological, and other phases of mental attainments, since these are the highest output of the earth's evolution. The latter seems to us the more comprehensive view.

(1) **The material effects of the mental element.**—Lyell long since urged that the direct work of man in changing the face of the earth was slight compared with that of the contemporaneous inorganic agencies. He called attention to the relative insignificance of the quarries, pits, cellars, and other excavations of man, compared with the work of streams, waves, and other inorganic agencies. There is justness in this view, but it needs qualification. It is to be observed that the mental era has but just begun, and that its effects are increasing with a rapidity quite phenomenal when measured by the slow pace of most geologic events. The excavations and transportations of material to-day show an enormous advance on those of Lyell's day, which was, geologically speaking, but a moment ago. The mile-tons of industrial freightage in the Mississippi basin are to-day not wholly incomparable with the drainage transportation of the same area a century ago. A century ago is named, because the surface was then covered with natural vegetation, and the normal effect of surface erosion, independent of man, was then experienced. At present the indirect effects of man's action are mingled with

those of natural processes, and these indirect effects are probably much more important than the direct ones. The removal of the native vegetation and the cultivation of the soil expose the surface to wash to a degree far beyond that prevalent when the surface was prairie sod, or leaf-carpeted forest, and denudation and transportation have been greatly multiplied in consequence. Not only has this cultivation increased the exposure to erosion, but, by increasing the rate of run-off, it has added to the erosive power of the streams. The ditching of swamps and other tracts of retarded drainage has contributed to this acceleration. The naked, soil-less uplands of some of the once populous kingdoms of the Orient, notably portions of Syria and Greece, are sad witnesses of the accelerated erosion that attends cultivation. The erosion of certain southern fields of the United States in the last forty years is another striking illustration. It is doubtful whether some parts of this region suffered as much erosion in the preceding five centuries as they have during the last one. On the other hand, some compensation is found in the reservoirs established for water-power, and in artificial devices for retarding and steadying stream flow.

In the light of considerations such as these, man may well be regarded not only as a potent geological agent, but as dangerously so to himself. The hope is that the intelligence that has wrought a change of surface conditions serviceable for the present, but dangerous to the future, will be so enlarged as to inspire a still more intelligent control of surface conditions which shall compass the future welfare as well as transient benefit.

Human modification of the animal and vegetal kingdoms.—Man's agency is also coming to be felt powerfully in the modification of the plant and animal life of the land and even to some extent of the sea. The larger animals that are not propagated by man are fast approaching extinction. At the present rate of extension of man's dominion, a century or so will see the disappearance of nearly every large mammal and reptile that he does not choose to protect or propagate. By way of compensation, certain selected animals are increasing and will doubtless continue to increase. The result is, therefore, likely to be a peculiar assemblage of animal life dependent strictly on the choice of a dominant type, a state of things that has apparently never occurred in n equal degree in the past history of the earth. How far the minor ns of life, especially the insect life, and the denizens of the sea, may be brought under this monopolistic control may not be predicted so easily.

A similar profound transition in vegetation is being forced by man. The native vegetation is rapidly being replaced by selected varieties, and by varieties that take advantage of conditions furnished by man. As the agricultural control of the earth becomes more complete and effective, a result toward which very rapid progress is being made, a new flora of man's selection will very generally prevail over the whole land surface of the globe. It is doubtful whether at any time in the history of the earth changes of flora and of fauna, and of surface, have been more rapid than those that are now taking place under the accelerating influence of man's action, and this accelerating influence springs not mainly from automatic or instinctive reaction, but from conscious impulse and intelligent direction.

(2) **The psychological factors as such.**—Are the introduction and the evolution of the psychological factors themselves to be regarded as subjects of geological study? We shall find that, at the outset, the geologic record is a complete blank so far as clear evidence of terrestrial organisms actuated by their own intelligence is concerned; that later, organisms with some apparent consciousness and intelligence appeared, and that the mental element increased apace unto its present attainment. We know that relationships of a sociological nature arose in apparent feebleness, and gradually evolved into more definite, higher, and more complex forms. By sociological factors we mean merely those conscious relations which one organism bears to another, of which the parental and the gregarious impulses are two fundamental expressions. For manifest reasons, the introduction and evolution of the psychological and sociological factors themselves have received little direct recognition as a portion of geological studies. The record of such factors in the fossils of past ages is necessarily obscure and imperfect, and the interpretation of what there is lacks certainty and precision. None the less, this psychological record, with all its imperfections, is beyond valuation, and must, we think, come to be an indispensable factor in the study of psychological and sociological evolution, for it shows, what nothing else can show equally well, the extremely prolonged history of that evolution, and it gives hints of modes and means which no study of existing stages can equally reveal. The organization of the Cambrian trilobites, for example, implies no small development of the senses

and of the coordinating faculties even at that early stage, and a study of the relations of these to their fellow creatures opens up the first known chapter in the sociological record of the earth's inhabitants. From this stage onward the progress in the development of the higher faculties, and of the sociological relations of the leading forms, is one of the most instructive phases of the great history. Such a study reveals the fact that many questions, narrowly supposed to be purely human, have had their prototypes in the earlier experiences of the animal kingdom. Some of these questions have found solutions, temporary or permanent, which passed under the test of ages to whose length human experience affords no parallel, and have received the sanction or disapproval of such tests according as they were well or ill adapted to the actual conditions involved. If one seeks the lessons of history in the largest sense, he cannot wisely neglect the prolonged record of the great biological family.

II. SPECIAL CONTRIBUTIONS OF THE ORGANIC KINGDOMS.

An essential part of the historical chapters of the second volume will consist of the description and illustration of the life progress of the successive periods. It will suffice here to give a preliminary synopsis of the kinds of record made by the several groups of plants and animals.

A. Contributions of the Plant Kingdom.[1]

The record of plants in the early geological ages is extremely imperfect. In the very earliest times the conditions seem to have been wholly unsuited to the preservation of any relics of life; but even after animal remains were abundantly preserved in the sea sediments, the plant record was still very meager for a long period. This was probably due in the main to two chief causes: (1) the probable softness and perishability of the early types of vegetation, and (2) the fact that vegetation is preponderantly terrestrial. At no time has marine vegetation reached a high development. Land conditions favor decomposition, transportation, and erosion, and through these, destruction; and only under rather occasional and exceptional conditions did the old lands leave a good record of their life. Nevertheless all the great groups of plants, viz. the Thallophytes (algæ, fungi), the Bryophytes (mosses, liverworts), the Pteridophytes (ferns, horsetails, lycopods), and the Spermatophytes (gymnosperms, angiosperms) have left some record.

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[1] Reference works: Scott, Studies in Fossil Plants, 1900; Zeiller, Éléments de Paléobotanique, 1900; Potonié, Lehrbuch der Pflanzenpaleontologie, 1899; Seward, Fossil Plants, 1898; Solms-Laubach, Fossil Botany, 1887.

REFERENCE TABLE OF THE PRINCIPAL GROUPS OF PLANTS.

Thallophytes (Thallus plants)	Algæ and algoid forms	Cyanophyceæ, blue-green algæ.	
		Chlorophyceæ, green algæ.	
		Rhodophyceæ, red algæ.	
		Phæophyceæ, brown algæ.	
		Diatomaceæ, diatoms.	
		Coccospheres Rhabdospheres	Pelagic algæ(?).
		Charophyta, stoneworts.	
	Fungi and fungoid forms	Phycomycetes, algæ-fungi, water-molds.	
		Ascomycetes, ascus-fungi, mildews.	
		Basidiomycetes, basidium-fungi, mushrooms.	
		Æcidiomycetes, æcidium-fungi, "rusts."	
		Schizomycetes, "fission-fungi," bacteria.	
		Myxomycetes, "animal fungi," slime-molds.	
	Lichens	Symbiont algæ and fungi.	
Bryophytes (Moss plants)	Hepaticæ, liverworts.		
	Musci, mosses.		
Pteridophytes (Fern plants)	Filicales	Filices, true ferns.	
		Cycadofilices, cycad-ferns.	
	Equisetales	Equisetæ, scouring-rushes, horsetails.	
		Calamites.	
	Sphenophyllales.		
	Lycopodiales	Lycopodiaceæ, club-mosses.	
		Lepidodendra.	
		Sigillaria and stigmaria.	
Spermatophytes (Seed plants)	Gymnospermæ (Naked seed)	Cordaiteæ, cordaites.	
		Cycadales (cycads)	Bennettiteæ. Cycadaceæ.
		Coniferæ, evergreens.	
		Ginkgoaceæ, ginkgo.	
	Angiospermæ (Covered seed) (Flowering plants)	Monocotyledoneæ, cereals, grasses, etc. (one-leafed seed).	
		Dicotyledoneæ, oaks, poplars, peas, etc. (two-leafed seed).	

The contribution of the Thallophytes (algæ, fungi, bacteria).—The Thallophytes embrace the simplest types of plants, and are probably the nearest present representatives of the ancestral forms. Some of them are minute one-celled organisms, as simple as an organism can well be conceived to be. The simple blue-green algæ of our fresh waters well represent this class. The most are, however, multicellular, and some (as the great seaweeds) rise to a degree of complexity and of a bodily segmentation resembling that of the higher plants. The various species are adapted to an extremely wide range of conditions; some live in hot springs at 170° Fahr., and some in Arctic seas at the freezing-point; some flourish in fresh water, some in brackish, some in salt water, and some even out of the water. This wide adaptation implies an ancient and plastic type. The fact that they flourish in waters so hot and sometimes also so sulphurous as to be fatal to most plants, suggests the possibility of their introduction during the very early volcanic stages of the earth, while conditions were yet uncongenial for other plants.

The geologic work of the thermal algæ is well shown in the beautiful travertine and sinter deposits of the Yellowstone Park (Figs. 215 and 218). At the Mammoth Hot Springs the deposits are calcareous, while at most of the other hot springs silicious deposits are formed, in both cases partly, but not wholly, by the aid of algæ. The beautiful yellows, reds, browns, and greens of these springs are not mineral coloring, but living plants.[1] In the calcareous waters, the algæ are believed to cause the deposition of calcium carbonate from calcium bicarbonate by consuming the second equivalent of carbon dioxide that rendered the carbonate soluble.[2] In the silicious waters, the process of deposition is not understood. Similar deposits by the aid of algæ take place in the geyser regions of Iceland and of New Zealand, in the hot springs of Carlsbad, where they have been well studied by Cohn,[2] and in most other hot springs. The same, or very similar, forms of algæ abound in nearly all waters, fresh and salt, but the question whether they make calcareous and silicious deposits in notable quantity appears not to have received as yet the critical investigation its importance deserves, except in a few special cases. It is clear, however, that in the cool waters such deposits do not reach the conspicuous amounts that they attain in the thermal springs. In the shallow waters of the ocean, especially in the warmer regions, lime-secreting algæ are abundant and make large contributions to the lime deposits.

Among the higher algæ are the lime-secreting corallines or nullipores (Rhodophyceæ, red algæ), once regarded as animals, which contribute a notable part of the calcareous substance of coral reefs. They are important geologic agents in the temperate and tropical seas, and have been traced as far back in time as the early Paleozoic era.

The Challenger reports[3] describe two forms of minute calcareous spherical organisms, Rhabdospheres and Coccospheres, as very abundant in the surface-waters of the temperate and tropical seas, and as important in contributing to the calcareous deposits of the sea-bottoms. The affinities of these bodies are in doubt, but they are regarded by Murray as probably pelagic algæ.

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[1] Weed. Ninth Ann. Rept. U. S. Geol. Surv., 1887–88, pp. 613–76; also Bradley M. Davis. Science, Vol. VI, 1897, pp. 145–57.

[2] Cohn. Abhandl. Schles. Gesell. Naturwiss., Heft II, 1862.

[3] Deep Sea Deposits, p. 257.

The stoneworts (Characeæ), an aberrant group of algæ inhabiting fresh and brackish water, secrete notable quantities of calcium carbonate in and around their tissues, and the accumulation of these gives rise to marl or limestone. It has recently been urged that our so-called shell-marls are mainly due to Charæ,[1] the molluscan shells being incidental rather than essential constituents.

In very ancient and also in some of the later strata, there are limestones that do not carry any visible fossils, and their origin is, therefore, debatable. There are also not a few limestones that are made up of a fine-grained base through which are scattered molluscan shells, corals, etc., in a fine state of preservation. The condition of these fossils bears rather adversely on the view that shells, etc., have been powdered in sufficient numbers and to a sufficient degree to form the compact base. In all these cases the usual explanations leave something to be desired. It is worth considering whether low forms of plants may not be among the undemonstrated agents in forming these apparently unfossiliferous limestones or parts of limestones. The calcium carbonate deposited by the algæ is in minute and delicate form, and is usually crystalline while yet in the living tissues. It is, therefore, easily subject to comminution and to such further crystallization as would obscure the minute features that constitute the evidences of algal origin.

The more complex and conspicuous algæ, the seaweeds, have left impressions of their stems and fronds on the marine beds of most of the periods, but they are usually obscure. Seaweeds are perhaps the source of the vegetal matter in certain carbonaceous shales and limestones. As seaweeds extract bromine and iodine and certain metallic ingredients from the sea-water, some of the iodine and bromine springs issuing from ancient marine deposits, and certain ores, may owe their origin to ancient seaweeds.

Diatoms, minute plants of the Thallophyte group, secrete a delicate framework of silica which becomes a contribution to the silicious deposits. Diatoms have sometimes contributed the material for very considerable beds, such as those of the ooze-bogs now forming in the marshes of the geyser basins of the Yellowstone Park,[1] and the diatom oozes of the deep sea (Fig. 353, p. 425).

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[1] C. A. Davis. Jour. of Geol., Vol. IX, 1901, p. 491.

Fungi, for obvious reasons, have left but scant traces of themselves.

Bacteria are believed to be recognizable as far back as the Paleozoic era. They are now the chief agents in the decomposition of organic matter, and may be regarded as the prime enemies of the fossil record. It is probable that similar decomposition took place actively in the earliest ages, for otherwise the remains of the ancient organisms should be more abundant. There is hence a theoretical probability that bacteria flourished as far back as the stratigraphic record goes. Not unlikely they were originally simple algæ that turned from the primitive habit of making their own food, to living on other organisms or their remains, and in so doing lost their power of manufacturing chlorophyl and of using inorganic carbon compounds. Their remarkable adaptation to the most varied conditions, and their extraordinary ability to endure the greatest vicissitudes of environment, support the view that they are a very ancient and plastic form.

At present certain bacteria are important to higher vegetation because of their ability to use the free nitrogen of the atmosphere and to combine it into forms available for the higher plants. It is not improbable that they have subserved this important function through all the known ages. Some experiments seem to show that certain of the existing algæ have this power, and possibly the ancestral forms of plants possessed it. The bacteria, being a derived and not an original form, could not have performed the function for the first plants. It is possible, of course, that the inorganic supply of nitrogen compounds was sufficient for plant life at the outset.

The contribution of the Bryophytes (liverworts, mosses).—The mosses and liverworts have left no certain record of their work in the earlier and middle geologic eras, and, if they existed at all, their contributions were unimportant. Although low forms of plant life, they are not primitive ones, as they are characterized by a definite alternation of generations implying a considerable time antecedent to the attainment of their present forms; hence there are no very cogent theoretical reasons for assigning them a place in early geologic history, though their absence cannot be affirmed. Some botanists think the Pteridophytes were derived from some ancestral form of liverwort, which, if true, would require the presence of the latter in an early geologic period; but the negative geological evidence relative to their presence favors the alternative view that the Pteridophytes were derived from some form of the Thallophytes by an independent line. In recent times, certain of the mosses, especially the sphagnum mosses, have played a notable part in the formation of peat accumulations. For this, their habit of growing in bogs, and of dying below while they continue to grow above admirably fits them.

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[1] Weed. Ninth Ann. Rept. U. S. Geol. Surv., 1887–8.

The contribution of the Pteridophytes (ferns, horsetails, lycopods, Sphenophyllum).—The Pteridophytes include the most important fossil plants of the earlier and middle geologic eras. To them we owe chiefly the great carbonaceous deposits of the Coal Measures and probably most of the disseminated carbons of the early and middle eras; perhaps also much of the natural oil and gas. Their special work is so conspicuous that it will be noted at length in the chapters on the Devonian and Carboniferous periods, and hence may be passed here with brevity. The ferns, now known more for their beauty than their importance, are the representative type of the group, and are really a wonderful family, having preserved their characteristic leaf-forms with a persistence attained by no other group of plants. The Paleozoic ferns are recognizable as such by every one, irrespective of botanical knowledge; indeed it is the detection of the differences, rather than the resemblances, between the ancient and modern forms, that requires expert knowledge. This continuity shows that since their introduction the changes of climate have never been so great as to prevent their propagation, without radical modification, in some part of the globe, and this fact rather narrowly limits the range of surface temperatures, and of other climatic vicissitudes. The persistence of the Equisetæ (horsetails, scouring-rushes) and the lycopods (club-mosses) bears like testimony, as does the persistence of life in general; but the rather delicate ferns are perhaps more obviously significant than most organisms.

The contribution of the Spermatophytes (seed plants, including gymnosperms or "evergreens" and angiosperms or "flowering plants").—The angiosperms, the dominant group to-day, make their appearance in the record in the latter part of the Mesozoic era, and their contribution is, therefore, relatively modern. They contributed to the coals, lignites, oils, and organic gases of the late geological periods, as did the Pteridophytes in the earlier periods, the latter participating, however, in the late deposits. Perhaps the most important function of the Spermatophytes lay in their superior serviceability as food for the higher land animals, by virtue of their seeds, fruits, and foliage. Neither the Thallophytes, Bryophytes, nor Pteridophytes, nor all combined, approach the Spermatophytes in food value for the higher types of animal life, and it is doubtful whether the higher evolution of the land animals could have taken place without the previous introduction of the seed plants. It will be noted in the historical narrative that the great placental group of mammals came in and deployed with marvelous rapidity, as geological progress goes, soon after the Spermatophytes became the dominant form of vegetation.

Plant life terrestrial rather than marine.—It is to be noticed that the chief development of all the great groups of plants took place on the land, or in the land-waters, rather than in the sea. This is preeminently true of the higher types, and appears also to be true of even the Thallophytes, although the number of individual algæ and their total mass is very much greater in the sea than on the land and in the land-waters. But the fresh-water algæ appear to possess in a higher degree than the marine forms those plastic and germinal characters from which new forms spring, and are probably to be regarded as the parental type. These are facts to be pondered on, since it has been the current opinion of geologists that life arose in the sea and was propagated thence to the land. The alternative view that life developed primarily on the land and in the land-waters and migrated to the sea is not, however, without its support in the plant world, as we thus see, and the plant world was the primitive one; the dependent animal world necessarily followed its development. The hypothesis of a terrestrial origin of life throws a very suggestive cross-light on many geological problems, as will be seen later, and it may well be entertained as an alternative working hypothesis until the facts are more fully developed.

B. CONTRIBUTIONS OF THE ANIMAL KINGDOM.[1]

As already noted, animal life is dependent on the decomposition of matter organized by green plants, and the conversion of its potential energy into active forms. Animals are, therefore, dynamic rather than constructive agencies. Nevertheless they transform organic vegetal matter into organic animal matter, and this is sometimes really an advance in organization. The organized animal matter is subject to preservation in some small degree, though it usually perishes. Some contribution is, therefore, made to the organic deposits, chiefly in the form of hydrocarbons. It is the view of some geologists that the natural oils and gases have an animal origin in the main.

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[1] Reference books: Zittel's Text-book on Paleontology, translated and edited by Eastman; Williams' Geological Biology; Nicholson's Manual of Paleontology.

REFERENCE TABLE OF THE PRINCIPAL GROUPS OF ANIMALS.[1]

Protozoa (The simplest animals)		Rhizopoda	Foraminifera. Radiolaria.
		Flagellata Infusoria Gregarina	Unknown in fossil state.
Cœlenterata (Sponges, corals, jellyfishes)	Porifera	Spongiæ	Calcareous sponges. Silicious sponges.
	Cnidaria	Anthozoa, coral polyps. Hydrozoa, hydroids and medusæ.	
Echinodermata (Crinoids, star-fishes, sea-urchins)	Pelmatozoa	Cystoidea, cystids. Crinoidea, stone lilies. Blastoidea, blastids.	
	Asterozoa	Ophiuroidea, brittle-stars Asteroidea, starfishes.	
	Echinozoa	Echinoidea, sea-urchins. Holothuroidea, sea-cucumbers.	
Vermes (Worms)		Platyhelminthes Rotifera Nemathelminthes Gephyrea	Rare as fossils.
		Annelida, sea-worms.	
Molluscoidea (Mollusc-like forms)		Bryozoa, sea-mosses. Brachiopoda, lamp-shells.	
Mollusca (Molluscs)		Pelecypoda, lamellibranchs, bivalves. Scaphopoda, tusk-shells. Amphineura, chiton. Gastropoda, univalves, snails, etc. Cephalopoda, nautilus, cuttlefish.	
Arthropoda (The articulates)	Branchiata	Crustacea.	Trilobita, trilobites. Gigantostraca, horse-shoe crabs. Entomostraca, ostracoids, barnacles. Malacostraca, lobsters, crabs.
	Tracheata	Myriapoda, centipedes. Arachnoidea, spiders, scorpions. Insecta, insects.	
Vertebrata		Cyclostomata, lampreys.	
		Pisces (fishes)	Selachii, sharks. Holocephali, spook-fishes. Dipnoi, lung-fishes. Teleostomi, ganoids and teleosts. (common fishes).
		Amphibia, amphibians, batrachians. Reptilia, reptiles. Aves, birds.	
		Mammalia (mammals)	Prototheria, monotremes. Metatheria, marsupials. Eutheria, placentals.

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[1] After Zittel in the main.

As dynamic organisms animals have need for supporting- and working-frames, for protective covering or housing, and for offensive and defensive weapons, and these have been constructed chiefly out of inorganic matter, and subordinately of indurated organic matter. It is through these that animals have made their chief contribution to the material of the geologic record. Skeletons and other hard parts to give internal stiffness or firmness; shells, plates, indurated integuments, and various other forms of external protection; teeth, spines, horns, and other means of gathering and masticating food, and of attack and defense, contribute material to the deposits, and form a record of the life activities and of the physiographic environment. All of the eight groups of animals, viz. Protozoa, Cœlenterata, Echinodermata, Vermes, Molluscoidea, Mollusca, Arthropoda, and Vertebrata, have left some record, but it is in all cases a very imperfect one.

The contribution of the Protozoa.—The Protozoa are related to the animal kingdom much as the Thallophytes are to the vegetable, and the two bear a close structural resemblance to one another. So near, indeed, do the Protozoa and the Thallophytes approach one another in their minuteness and simplicity, that the place of not a few organisms is in doubt, and the two kingdoms, in general so different, seem here to blend in the group Flagellata. The Protozoa are usually very minute one-celled organisms with very little differentiation of tissue or organs. Of the four classes of Protozoa, only one, the Rhizopoda, is found in the fossil state. The rhizopods secrete silicious skeletons, and calcareous, silicious, and chitinous tests of a great variety of forms, and this gives them geologic importance. The deep-sea oozes and the chalk deposits are their best-known contributions at present. They have probably played a more important rôle in the formation of ordinary limestones and silicious silts than can be demonstrated, because of the delicacy of their relics and the ease with which these are pulverized by wave-action in the shallow seas, or changed by recrystallization or by concretionary aggregation. The globigerina oozes are formed largely from the calcareous shells of Foraminifera (Fig. 351), one of the orders of rhizopods, among which the genus Globigerina is a leading form. Those forms which make the deep-sea oozes live, not on the bottom, but near the surface of the open sea, and on the death of the organisms, the shells, tests, and skeletons sink to the bottom. Chalk is formed in a similar way from calcareous Foraminifera, but not necessarily in very deep water. Foraminifera live in shallow water as well as in the open sea, and in this case they sometimes creep on the bottom or are attached to algæ, but their deposits in shallow water are usually much obscured by other kinds of deposition and by destructive action. Some of the foraminiferal shells are divided into chambers and assume various spiral forms, of which the *Nummulites*, named from their resemblance to coins, are notable examples. These formed an important part of the nummulitic limestone of the Eocene period.

The radiolarian ooze is characterized by the silicious tests of various members of the silica-bearing order, Radiolaria. The "Barbadoes earth" and "Tripoli" are notable deposits of fossil radiolarians.

The contribution of the Cœlenterata.—The Cœlenterata embrace the sponges, the coral polyps (Anthozoa), and the hydroids and medusæ (Hydrozoa). The contribution of coral polyps to the formation of limestone is most important, and is too familiar to require elaboration here. The corals range throughout nearly the whole fossiliferous series, and their development will be followed and illustrated in the historical chapters.

The sponges are widely represented by their spicules, and not uncommonly their aggregate form is preserved even in very ancient strata. Their contribution is largely silicious, but is partly calcareous. The hydroids and medusæ have left little trace of themselves in the rocks, although impressions supposed to represent medusæ are found in strata as early as the Cambrian. Certain coral-like forms, as the Millepores, Tubularia, and Stromatopora, are classed as Hydrozoa. The graptolites, delicate leaf-like floating forms, very serviceable in marking exact horizons on different continents because of their free distribution, are also classed here.

The contribution of the Echinodermata.—Under the echinoderms are grouped the crinoids (sea-lilies), cystoids, blastoids, ophiuroids (brittle stars), asteroids (starfishes), echinoids (sea-urchins), and holothuroids (sea-cucumbers). This is one of the marked groups of ancient as well as modern life, and its beautiful fossils grace every period in which life relics are well preserved. The cystoids and crinoids, and later the blastoids, were prominent in the Paleozoic ages, while the remaining forms were more conspicuous later, though early introduced. All divisions, except the holothuroids, whose softness prevented, have left a good record, as fossil records go. Their relics are chiefly calcareous, and they most abound in the limestones, some of which are largely made up of their remains, as the encrinital limestone (Fig. 349). They will be subjects of frequent comment and illustration in the historical chapters.

The contribution of the Vermes.—Most of the worms are ill adapted to fossilization and are not known in the fossil form. The segmental worms of the sea, the annelids, however, left some traces of themselves in tubes and borings and in tracks and sometimes by fossil jaws and teeth. They range from the earliest fossil-marked horizons onward, but seem to have always been an inferior group.

The contribution of the Molluscoidea.—This group includes the bryozoans, whose fossil products closely resemble the minute-celled corals, and the brachiopods, whose shells closely resemble those of the molluscs. Both are calcareous and make important contributions to the formation of limestone (Fig. 350). A few brachiopods secrete calcium phosphate instead of calcium carbonate. Both classes have a great geologic range and their fossils are valuable aids in identifying and correlating formations. Probably the brachiopods are more utilized for this purpose than any other single class. They are the symbol of conservatism and persistence, ranging from the Cambrian to the present time, and embracing some forms that have scarcely changed to the extent of generic difference in that time.

The contribution of the Mollusca.—The molluscs have also ranged from the earliest well-recorded times, and some divisions, as the pelecypods (lamellibranchs, embracing clams, oysters, etc.) and gastropods (snails, etc.), have undergone no very marked change beyond a rather ample and progressive development; but others, as the cephalopods (nautilus, squids, cuttle-fish, etc.), mark out the progress of the ages by distinct and striking changes of form. Their shells are chiefly calcareous and they have contributed materially to the formation of limestone. Muddy and sandy bottoms are, however, more congenial to the pelecypods and gastropods than to the corals, crinoids, and many other limestone-forming types, and hence fossils of these molluscs frequently abound in shales and sandstones and give them a calcareous element. In sandstones, however, the calcareous matter is often dissolved out and only the casts of the shells remain. The molluscs will be much cited and illustrated in the historical chapters.

The contribution of the Arthropoda.—This group embraces the crustaceans, myriopods, spiders, and insects. The hard parts of their bodies are mainly horny or chitinous forms of organic matter, and hence their relics differ notably from the inorganic calcareous and silicious remains of most of the preceding forms. The Arthropoda did not at any time form a notable stratum of rock. Their geologic value lies chiefly in what they teach of the progress of life and its relations, and the aid they render in correlation and identification. In these respects the group is a notable one. It was represented in the early fossiliferous strata by the trilobites, one of the most interesting of all types of fossils. These were probably the most highly developed organisms of their times and give the clearest hints of the stage of psychological and sociological development that had been reached when first the record of life is opened to us. The record of the myriopods, spiders, and insects dates from the middle Paleozoic, and gives the first clear hints of animal life on the land.

The contribution of the Vertebrata.—In the vertebrates the dynamic or working organism may be said to reach its highest expression, unless it be in the flying insects, and their inorganic residue becomes relatively unimportant in rock formation. Although the greatest of all animal types in most respects, it has never formed more than trivial beds of rocks. There are occasional "bone beds," but they are thin and limited in extent, and only partially formed of vertebrate matter. The geological importance of the vertebrates lies in the higher field of life evolution and in its mental accompaniment. Fishes excepted, the vertebrates are mainly land types, and have for their chief colleagues plants and insects. The other groups of animals are mainly, though not wholly, marine. The vertebrates have little place in the Paleozoic record, except near its close, but they dominate the Mesozoic and Cenozoic eras, and are conspicuously the master type to day.

III. THE ASSOCIATIONS AND ECOLOGICAL RELATIONS OF LIFE.

A. The Basis of Floras and Faunas.

Geologic interest is not confined to the kinds of plants and animals that have lived and the contributions they have made to the deposits, but embraces also their assemblage into floras and faunas, and the relations of these assemblages to the prevailing physiographic features. These assemblages and relationships are among the most suggestive factors of the earth's evolution, and are the most instructive for purposes of comparison with human history, and for forecasting the future of man and of the whole biological kingdom. Moreover, floras and faunas, as such, are used in the correlation of formations, and in this application they give surer results than correlations by individual species. A particular species may live far beyond the usual period of a species, and if fossilized in one region in its early history and in another in its late history, the two formations might be referred erroneously to the same stage. This is far less likely to happen with a whole assemblage of forms. There is a similar liability to error in interpreting migrations on the basis of a single or a few species, for a single species or a few species may be transported by unusual or accidental means, so to speak, when there is no normal pathway for general migration, and when no systematic migration takes place. In most of the great questions that arise concerning the connections and disseverances of the continents, and concerning the unions and separations of the oceans, which are the fundamental causes of the migrations and of the isolations of plants and animals, typical floras and faunas are to be studied, rather than isolated species or sporadic forms. A brief sketch of the leading causes and consequences of these special assemblages of plants and animals may aid in appreciating the underlying significance of floras and faunas, and in interpreting their meaning as they are met in the study of the strata. A part of these grow out of the relations of the organisms to one another, and a part out of the relations of the organisms to their environment.

(1) *Assemblages Influenced by the Mutual Relations of Organisms.*

(*a*) **Food relations.**—The relations of food-supply are among the most obvious reasons for assemblages. As animals are dependent directly or indirectly on plants for their food, they must gather where the plants grow, or in the currents in which the plant products are borne. Whatever determines an assemblage of plants also causes, or at least invites, an assemblage of animals. Whatever causes an assemblage of particular plants, invites an assemblage of the particular animals that use these plants. Animals that feed on plants are in turn preyed upon by other animals, and these in turn by others. A whole train of organisms may, therefore, be gathered into a region by the conditions that foster a certain kind of vegetation there. In interpreting the physical significance of such a train, it is obvious that the head of the train carries the fundamental meaning. The dependent creatures that follow the primary forms may be only incidentally, and perhaps very slightly, adapted to the physical environment.

(*b*) **Adaptive relations.**—Organisms depending on other organisms for food or other necessary conditions of life, present many forms of adaptation the better to secure their food and to use it. These adaptations are the consequences and the signs of the assemblage, and are of the greatest service in interpreting the place and significance of the organisms in the assemblage. Teeth usually reveal the food of their possessors, and hence teeth are among the most significant of fossils. Fortunately their functions require them to be hard and durable, and hence well suited to fossilization. The growth of low plants into trees forced a notable series of adaptations in the animals that fed upon them in the matter of height, of reaching members, of climbing, and probably at length of parachuting and flying. In these and similar ways the floras and faunas took on special phases because of the mutual relations of their members.

(*c*) **Competitive relations.**—The assembling of plants and animals, with their prodigious possibilities of multiplication, brought competition, and with it a struggle for food which often became a struggle for existence, and out of this grew innumerable modifications of form and habit. These have become so familiar since the great awakening caused by the doctrines of Darwin and Wallace that they need no elaboration here.

(*d*) **Offensive and defensive relations.**—Within limits, plants are benefited by the feeding of animals and respond by developing seeds and fruits that especially invite such action, their compensation being found in planting and distribution. It is obvious that, on the whole, the continued growth of plants is largely dependent on the renewal of a supply of carbon dioxide through the agency of animals and some plants, bacteria in particular. Otherwise the supply would become so reduced as to greatly limit plant life. It has been estimated[1] that the whole of the present supply of carbon dioxide would be consumed by plants in one hundred years if the consumption continued at the present rate and no carbon dioxide was returned. It is now well known that the so-called decay by which carbon dioxide is freed is due more to microscopic organisms than to inorganic processes. It seems clear, therefore, that the continued activity of plants is largely due to their consumption by animals and other plants. But still, though the larger good of plants is conserved by the predaceous action of animals, and of certain parasitic and saprophytic plants, their individual preservation is often conserved by defensive devices, such as thorns, poisons, bitter compounds, etc. This is notably true in desert regions where the conditions are hard and the total extinction of plants would be threatened if animals were permitted to feed freely upon them. Within the animal world, the preying of one form upon another is the main source of that great struggle for existence which has characterized the whole known history of life, and has been one of the influential factors in shaping the evolution of life and in modifying the special aspects assumed by the floras and faunas of each period.

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[1] S. W. Johnson, How Crops Feed, p. 47.

Implied forms of life.—The full meaning of the fossils of any period can only be gathered by duly considering these relationships in their interpretation. The existence of animals implies the existence of plants in supporting abundance, whether the record contains their relics or not; an animal with a protective covering implies an enemy; a tooth of a specific kind implies the appropriate class of food, etc. While inferences of this kind are subject to error, they are at present the only means by which the faunas and floras of most ages can be rounded out into a rational assemblage of organisms, that is, an assemblage that affords the necessary food for its members and an adequate function for the offensive and defensive devices which its members present. Only a small part of the life that lived was fossilized, and only a small part of the fossils actually carried in the strata have been collected, because only a small part of the strata are exposed at the surface. The direct record now accessible is, therefore, very incomplete and hence the need—and in the need the excuse—for adding the forms that are implied by the character of the known fossils.

(2) *Assemblages Influenced by Environment.*

It has been noted that some animals depend for existence on other animals; that ultimately all animals depend on plants, and that green plants alone can make food directly from inorganic material. Green plants, therefore, head the train of dependencies, and their relations to the physical conditions that surround them are the primal relations.

Plant societies.[1]—The control of physical conditions has been sufficient to develop special associations or societies of plants by fostering those adapted to these conditions and eliminating those that are not. Among these are (1) the hydrophytes ("water plants"), embracing those that grow in water or in very wet situations; (2) xerophytes ("drought plants"), embracing the opposite class, which are adapted to very dry situations; (3) mesophytes, including those suited to conditions lying between these extremes, the great middle class to which the prevailing upland vegetation belongs; and (4) the halophytes ("salt plants"), which are dependent on the presence of certain salts, and embrace such plants as are found on the seacoast, around salt springs, on alkaline flats, etc. The characters which distinguish the xerophytes from the hydrophytes and mesophytes have special geological interest, as they aid in determining the climatic conditions, a feature whose interest increases as the variability of the ancient climates is more fully recognized.

Within these greater groups there are special minor associations determined by soil, temperature, topography, subjacent strata, and by the relations of the plants to one another.[2] These natural groups are valuable indications of the agricultural capabilities of the districts occupied by them. They may be regarded as the outcome of Nature's experiments in crop-raising, running consecutively through thousands of years. They are natural correlations of compatible members into communities of plants. Some members of the society are obviously dependent on others, as certain forms of undergrowth on the shadowing of the upper growth, as of vines upon supporting-trees, etc. There is probably a more occult relation in some cases, the effects of certain plants on the soil being sometimes advantageous to other plants, and sometimes harmful, as illustrated in the conditions that require a rotation of crops.

The chief point of geologic interest lies in the fact that floras are not mere miscellaneous mixtures of plants that happen to live in a given area at a given period, but are organized communities, in a more or less definite sense. They therefore imply more or less definitely the physical conditions which are congenial to them, and thus furnish the basis for interpreting such conditions in the past, so far as the floras are well preserved. The faunas, especially the land faunas, being primarily dependent on the floras, furnish a basis for interpretations of like import.

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[1] Reference works: Plant Relations, Coulter, 1900,—a convenient elementary work; Schimper, Pflanzengeographie, 1898; Warming, Lehrbuch der oekologischen Pflanzengeographie, 1896; Cowles, Botanical Gazette, Vol. XXVII, 1898.

[2] One of the earliest attempts to map these and develop their significance and value is found in Vol. II, Geol. of Wis., 1873–77, Native Vegetation, pp. 176–87.

B. THE INFLUENCE OF GEOGRAPHIC CONDITIONS ON THE EVOLUTION OF FLORAS AND FAUNAS.

The geographic features of the earth impose on organisms a complex series of influences which modify the evolution of life and produce faunal and floral variation on a large scale. The larger assemblages of life, which inhabit a continent or dwell in a great sea, are designated faunas and floras, as well as the smaller assemblages just discussed, but obviously in a broader and in a different sense. The disseverance of the land by the sea, or of the sea by the land, isolates the life and forces independent development. The introduction of cold zones, desert tracts, or other potent climatic belts has somewhat the same effect. So, measurably, does the raising of a mountain range or a plateau, or the sinking of critical portions of the sea-bottom.

The development of provincial and cosmopolitan faunas.[1]—If a region is isolated from other regions by the cutting off of all ready means of intermigration, as by the formation of an island from what had been a peninsula, or of an inland sea from what had been a bay, the flora and fauna are developed by themselves without much influx of other forms, and hence become local or provincial. This is usually more marked in the case of the fauna than of the flora, because the latter has more ample means of dispersion, on the whole, and so the fauna may for convenience be taken as the type. A good illustration is the native fauna of Australia which was once connected with Asia, but has long been separated from it. Previous to importations by man, this continent had a very peculiar and distinct fauna, descended from its Mesozoic inhabitants. Most of the isolated islands have peculiar faunas, but in many cases they were isolated from the beginning, having been built up by volcanic action from the bottom of the sea, and their faunas are due to the accidents of transportation and to the development of these sporadic forms in isolation.[2]

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[1] Chamberlin. A Systematic Source of Evolution of Provincial Faunas, Jour. of Geol., Vol. VI, 1898, pp. 597–609.

[2] Wallace. Island Life.

It is evident that whenever any geographic change introduces a barrier to migration, the faunas of the dissevered portions will, in all probability, develop along different lines, and will diverge into provincial faunas. On the other hand, any geographic change that unites areas and leads to intermigration, tends to a community of fauna or to cosmopolitanism. These tendencies have been markedly felt all through the geologic ages, and constitute one of the most vital features of their history. When continents are connected, their faunas intermingle and the exchange gives rise to common forms. They tend to blend into one great fauna except so far as the local differences develop those minor assemblages previously discussed. When continents are separated, they tend to develop peculiar faunas, as do islands, but on a larger scale. This is very obvious in the case of the land life, but needs more special statement for the oceans.

The oceans constitute a single body of water with ample connections and stirred by a system of constant circulation. Probably this has been true for most of known geologic time. A single cosmopolitan fauna of the largest type might be expected. This is in a measure realized in the pelagic fauna of the open ocean, though this is somewhat modified by the climatic zones. But the marine faunas that are fossilized in the known strata, and have most geologic interest, are, with rare exceptions, not those of the open ocean, but those of the shore zones and of the shallow seas. Now, although these shore belts and shallow seas are broadly connected with the great ocean body, and are usually regarded as a part of it, they are singularly separated from it, or rather they are singularly separated by it, so far as the life dependent on shallow-water conditions is concerned. To this life, the deep sea is a barrier not quite as effective as the land, but still a barrier. The key to this important fact may be found in a consideration of *the vertical distribution of life.*

The great horizon of life is at or near the contact zone of the atmosphere with the hydrosphere and lithosphere. Life declines with increasing altitude, partly because of the lowering temperature, and partly because of the increasing tenuity of the atmosphere. The successive changes of plant and animal life with the ascent of mountains and plateaus is familiar. Life declines in descent into the sea chiefly from lack of light, and secondarily from the lowering of temperature. Light is essential to the formation of chlorophyl and, through it, of all other organic compounds. The chlorophyl-forming plants are, therefore, limited to such depths as are penetrated by the rays necessary for the photosynthesis of organic matter. Vision is cut off within 200 to 300 feet, and most plant growth takes place above that depth. Photographic effects become feeble or inappreciable at 1000 to 1200 feet.[1] The photosynthesis of plants is chiefly aided by the lower and middle part of the spectrum, while the ordinary photographic work is chiefly done by the upper end, so that the photographic limit is below the photosynthetic limit. Microscopic plants are sometimes found lower than these limits, but they may have been carried below their working limits by currents or other incidental agencies. For all general purposes, the limiting depth of living carbon-compounding plants may be set at 100 fathoms, as a generous figure—about the average depth of the border of the continental shelf—while the vast majority flourish only in the upper third of this depth.

Life does not cease here, for the products of this surface-life sink to greater depths and are fed upon by forms of sea animals that have become adapted to the dark and cold abyss of the ocean. Obviously, these deep-sea forms are a very distinct type of life, and constitute a fauna of the most pronounced kind, the abysmal fauna. Another distinct fauna occupies the open-ocean surface, the pelagic fauna. Still a third fauna occupies the shallow-water tract, whose bottom lies within the light zone—*the photobathic zone*—and embraces the animals that are dependent on the plants of this zone, or on its light and warmth, and that are more or less fixed to the bottom or confined to the zone because their food is there.

The physical plane of demarkation between the surface or pelagic fauna and the abysmal fauna is much more distinct and more fundamental than any that is found in ascending above the surface of the sea. The habitat of the shallow-water fauna is limited below by the darkness, limited above by the water-surface, limited at one side by the land, and limited on the other side by the deep sea. It is hemmed in vertically between two planes only a few hundred feet apart. Laterally, it is confined to a narrow belt about the borders of the continents and to the more or less land-girt epicontinental seas. Its vertical limits are fixed; but its lateral extent varies with the relations of the sea to the surface of the continental platforms.

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[1] For data, see Walther's Einleitung in die Geologie, pp. 35–45.

This variation profoundly affects the development of the fauna. When a major deformation of the earth takes place which increases the capacity of the oceanic basins, the water is drawn down into them more fully, and correspondingly retreats from the continental shelf. The shore is thus carried out toward or to the border of the shelf, or even perhaps down to some line on the abysmal slope. In either case, the zone of shallow water suited to the photobathic life is narrowed, and at points it may be practically cut in two. There are, however, shelves and tracts that were below the light zone before, which now are brought within it by the lowering of the sea-level. Into these, as into harbors of refuge, the life migrates so far as it may. But these tracts are less prevalent and continuous than the typical continental shelf, and under the conditions supposed they would be but imperfectly connected with each other by available shallow-water tracts. (The steep shelving shore tracts, although furnishing a shallow-water connection possibly available for some species, would be unsuited to others and, under certain conditions of the sea-currents, would be an effective barrier.) To these limited tracts, therefore, the life of the photobathic type is restricted and measurably isolated, and develops into local and provincial faunas.

After a deforming movement has ceased, the seashore habitually advances, developing a new continental shelf, and in time new epicontinental gulfs and seas. In this it is assisted by the erosion of the continent and the filling of the sea, and probably by the slow settling of the continents. As the sea-shelf broadens, the isolated tracts, the harbors of refuge, become connected, and migration is facilitated. When the connection becomes general and broad, and when epicontinental seas have formed available tracts across the face of the continents, a general commingling of faunas follows, and a cosmopolitan fauna results.

In the same way, but more obviously, when the land is extended and connection between the continents becomes general, there is migration and commingling of the land faunas and floras, and cosmopolitan communities are the result.

It is obvious that the development on the land is the reciprocal of that in the sea. When the seas are extended and their life is tending toward cosmopolitanism, the lands are dissevered, and their life is tending toward provincialism, and *vice versa*. When, however, the land is greatly extended, it is usual [illegible] ntain ranges, and other products of the deformation which extended it, and these form barriers. Desert wastes and other inhospitable tracts, and even glaciation, are liable to develop as secondary consequences, and to interpose barriers, and hence the cosmopolitanism of the land-life is liable to be less complete than that of the sea-life.

Restrictive and expansional evolution.—It is obvious from the last discussion that if the picture of the earth's movement above drawn be true, the areas available for particular classes of life may vary greatly from age to age. At times the shallow-water sea-life may be forced to retreat into a very narrow tract on the border of the land, and into chance expansions here and there. In being crowded into this limited tract, perhaps also less adapted for a habitat on account of the change, the life is subjected to severe competition and to hard conditions, and must experience in an intensified degree the effects of the struggle for existence. Whatever of evolutionary potency there may be in such a struggle under such restrictive conditions should be revealed in the modifications of the fauna that ensued.

On the other hand, when the shallow seas are generally extending themselves upon the land and the land is being base-leveled, and thus adapted to shallow submergence, the shallow-water life enjoys an enlarging realm, and should reveal the effects of evolution under expansional conditions. In affording a comparison between these opposite and alternating phases of restrictional and expansional evolution, geology makes one of its great contributions to the external causes and conditions of organic evolution. These will come under repeated consideration in the historical chapters.

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